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UNCONSOLIDATED SEDIMENTARY AQUIFERS: REVIEW NO.10 - APPLICATIONS OF SURFACE AND AIRBORNE GEOPHYSICS

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Sigatoka River flood plain, Fiji

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Unconsolidated Sedimentary Aquifers

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UNCONSOLIDATED SEDIMENTARY AQUIFERS (UNSAs)

PREFACE

This Review is one of a set of reports prepared as part of a project entitled "Groundwater Development in Alluvial aquifers", Project No R5561 (BGS 93/2), under the ODA/BGS Technology Development and Research (TDR) Programme of aid to the developing countries. The project addresses all unconsolidated sedimentary aquifers (UNSAs) not only alluviums.

Each review comprises a detailed state of-the-art synthesis of a particular aspect of the hydrogeology of UNSAs and is intended to assist overseas professionals in the evaluation and development of groundwater resources in UNSAs. An overall Handbook will also be produced, in order to summarise these detailed reviews and draw together the results of the project.

This particular review describes the numerous applications of surface- and airborne geophysical techniques to aid the efficient exploitation of the groundwater contained in UNSAs.

This review is a compilation of existing knowledge. It is intended to be updated as appropriate, following the results of research which will be carried out during the lifetime of the project, which is scheduled to run until 1996. One such update has already been included; this covers the application in June 1995 of the most recent geophysical technique (Electro Kinetic Surveying) to the investigation of Vietnam's Red River alluviums.

The project is funded by ODA as part of their research and development programme designed to improve living standards and conditions in the world's developing countries.

Project Manager: Dr R Herbert
Hydrogeological Advisor to ODA
British Geological Survey

INTRODUCTION

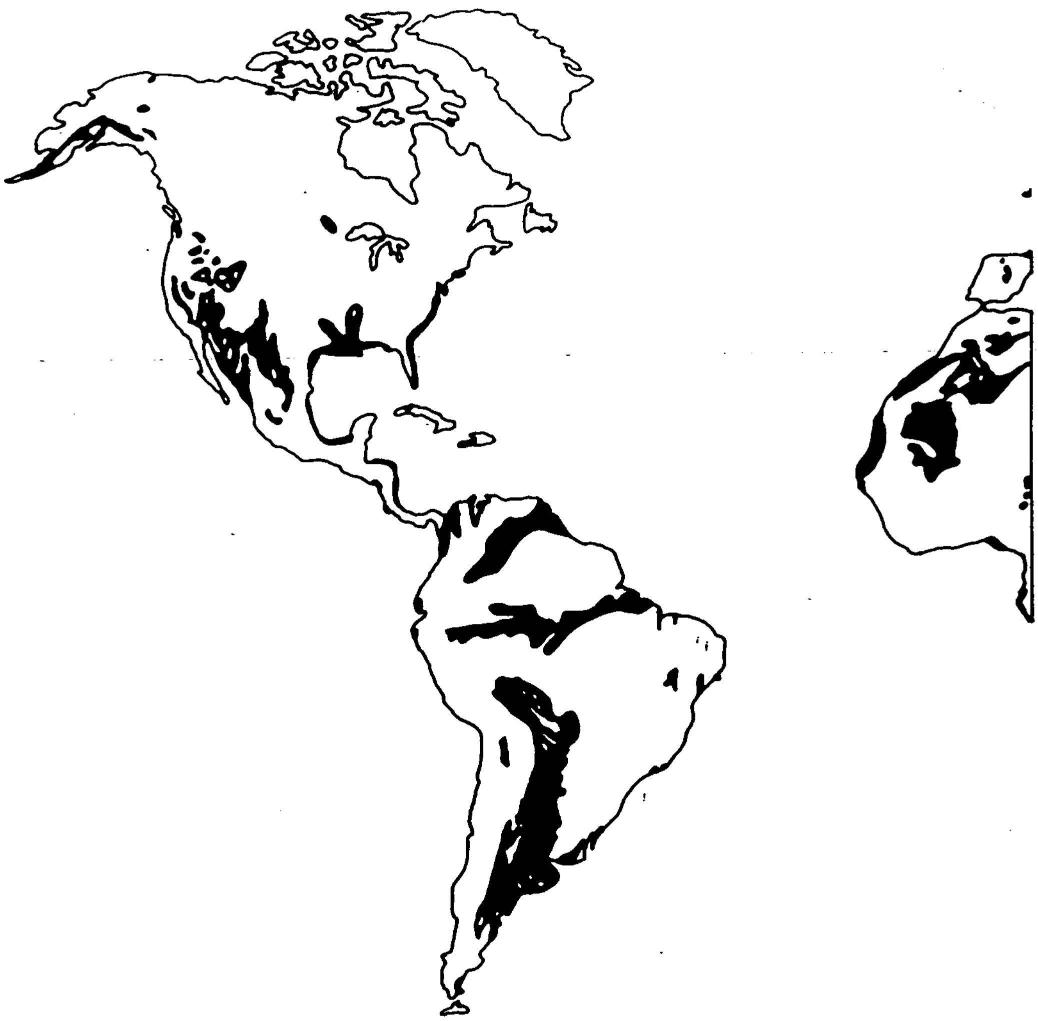
WHAT ARE UNSAs AND WHY IS IT IMPORTANT TO UNDERSTAND THEM?

UNSAs are unconsolidated sedimentary aquifers. These are the water-bearing strata within the swathes of unconsolidated sediment that mantle much of the earth's surface. There is no clear dividing line between UNSAs and aquifers in consolidated rocks, as lithification is a gradational process: deposits a hundred years old can be lithified, while some deposits 500 million years old are still essentially unlithified. However, for most purposes, UNSAs can be understood as deposits which have accumulated over the past few million years, that is during Quaternary and Neogene (late Tertiary) time. They are important sources of water in many parts of the world, and in particular constitute the only major sources of groundwater for vast areas throughout the developing world. In the influential text-book *Hydrogeology* by Davies and De Weist it says:

> "The search for ground water most commonly starts with an investigation of nonindurated sediments. There are sound reasons for this preference. First, the deposits are easy to drill or dig so that exploration is rapid and inexpensive. Second, the deposits are most likely to be found in valleys where ground-water levels are close to the surface and where, as a consequence, pumping lifts are small. Third, the deposits are commonly in a favourable location with respect to recharge from lakes and rivers. Fourth, nonindurated sediments have generally higher specific yields than other material. Fifth, and perhaps most important, permeabilities are much higher than other natural materials with the exception of some recent volcanic rocks and cavernous limestones."

To date, though, few attempts have been made to understand the detailed internal structure of unconsolidated aquifers even though such knowledge may be crucial to the long term success of any water development project. This shortcoming is probably the reason why the operational lives of many water boreholes are frequently much shorter than expected.

Understanding of the internal structure or "architecture" of many types of sedimentary deposit has, however, advanced greatly over the past couple of decades. Part of this research has been academic, but much has been sponsored by the oil industry, so as to better predict the possible location of oil within sedimentary traps. Oil, like water, is most profitably located within bodies of relatively coarse-grained and porous sediment. Thus, there is obvious scope for applying this recently gained understanding to hydrogeological problems. Advances have also been made in the understanding of the geometry of complex "soft-rock" deposits by the application of appropriate combinations of investigative techniques, including remote sending, rapid geophysical methods and new drilling techniques. The combination of these bodies of knowledge can provide a framework for locating and assessing UNSAs.



MAJOR AREAS OF UNCONSOLIDATED SEDIMENTARY AQUIFERS WORLDWIDE

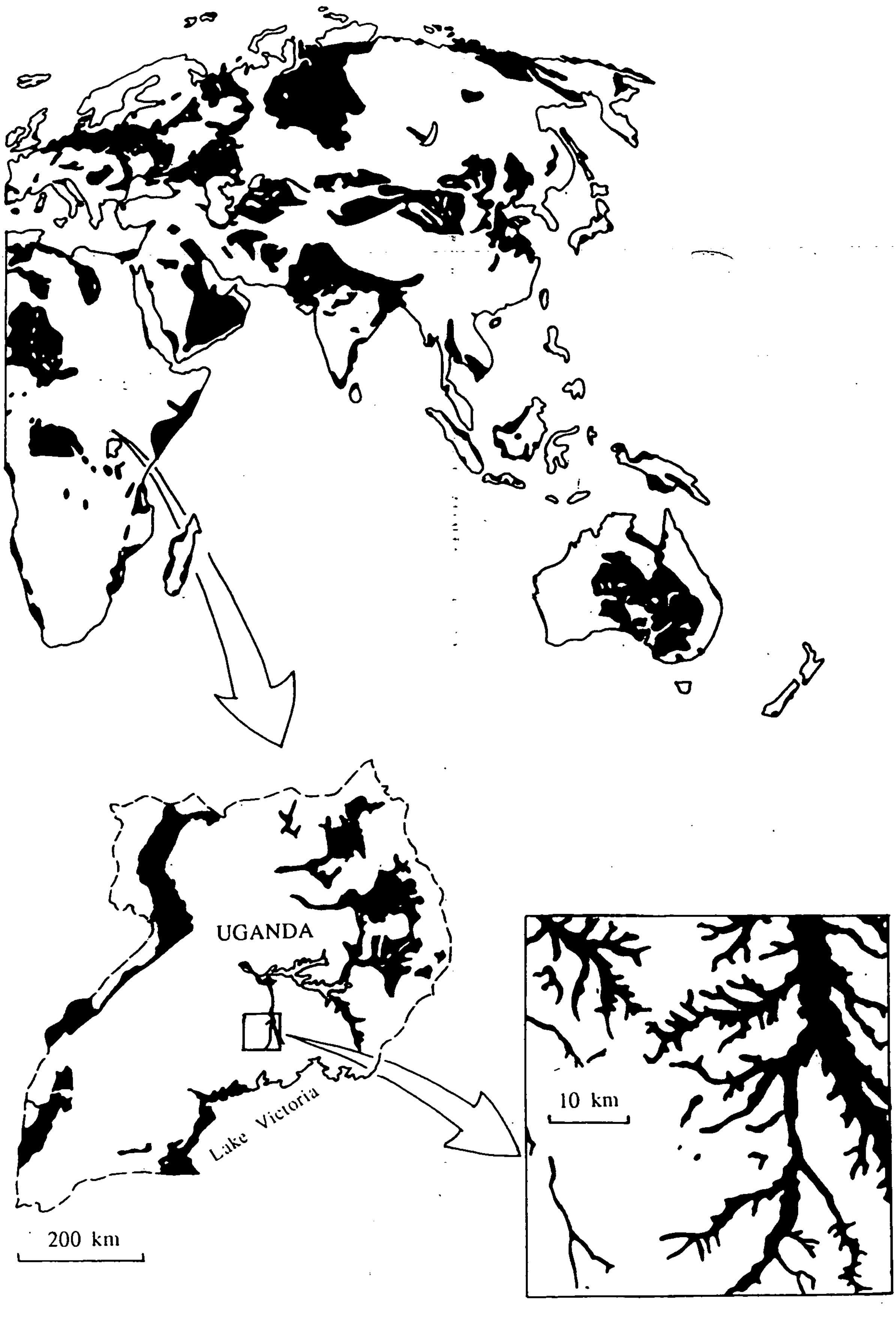
- The map shows the distribution of the thickest and most extensive Quaternary deposits in the world. The great majority of these are unconsolidated, and many include water-bearing deposits (UNSAs).

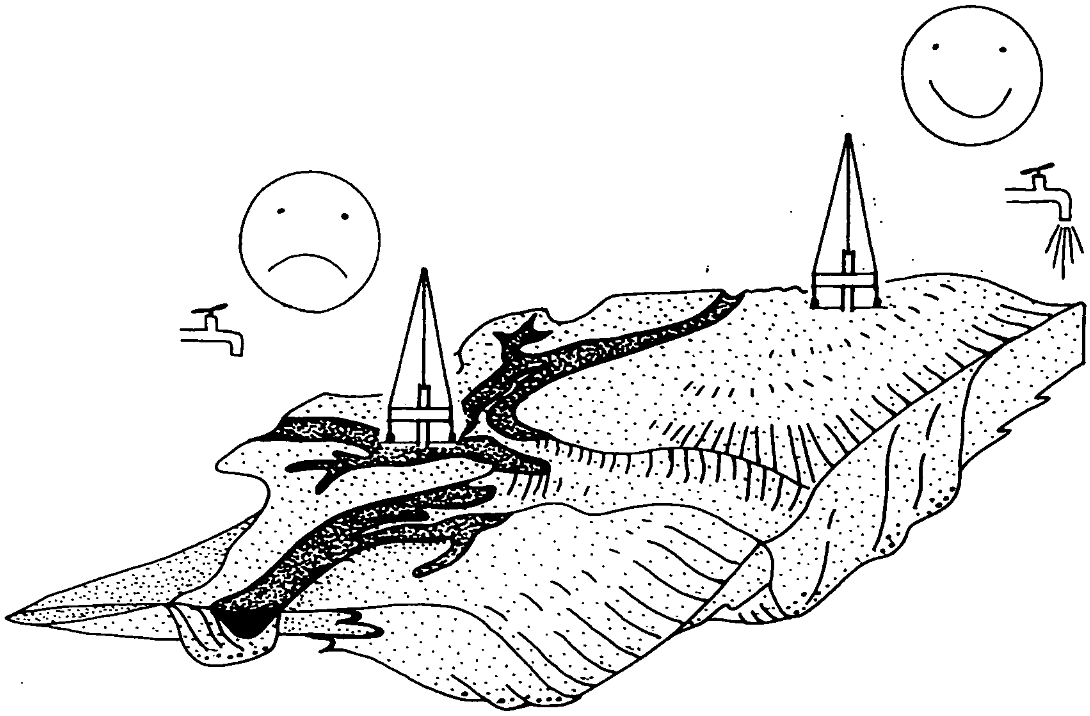
- A generalised world map such as this, though, severely under-estimates the true extent of UNSAs worldwide. This is because:

  - unconsolidated pre-Quaternary deposits are omitted; these too have a wide distribution, though are difficult to delineate (as they grade into consolidated deposits); they too can include significant UNSAs.

  - the simplification of linework necessary at this scale means that a large proportion of unconsolidated deposits have had to be omitted. The inset map shows the example of Uganda, which seems to have no unconsolidated sediments at the global scale, while significant and extensive deposits 'appear' once the country is looked at more closely. At a yet larger scale the unconsolidated sediments appear yet more widespread. The message is clear. *Unconsolidated sediments, and therefore UNSAs, are ubiquitous.*

Diagram data modified from various sources.


UGANDA
Lake Victoria
200 km
10 km



Sedimentary bodies are characterised by variably complex geometry and internal structure. These properties exert a strong internal control on the location, quantity and quality of groundwater. Diagram adapted from Galloway and Hobday (1983).

APPLICATION OF SURFACE AND AIRBORNE GEOPHYSICS TO GROUNDWATER EXPLORATION AND MANAGEMENT

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APPLICATION OF SURFACE AND AIRBORNE GEOPHYSICS TO GROUNDWATER EXPLORATION AND MANAGEMENT

1. AIMS OF THE REVIEW

This review outlines the physical principles of relevant geophysical techniques and then describes numerous applications of surface and (less commonly) airborne methods to aid the efficient exploitation of unconsolidated aquifers. The aims of the review are:

- to provide an overview of the surface and airborne geophysical techniques that can be applied in the study of unconsolidated materials,

- to highlight the advantages and limitations of the techniques in the various applications and to summarise these findings for quick reference,

- to provide key references where the specific applications are described in detail.

1.1 Background

Surface geophysical techniques are non-invasive and cost-effective and can indicate the optimum site for a monitoring- or production well or borehole at a fraction of the cost of a programme of exploratory drilling. Typical applications include the determination of the thickness of unconsolidated deposits and the nature (grain size, clay content etc) of these deposits. Several techniques can be used to detect the depth of the water table and hence permit estimates to be made of the required drilling depths and the total volume of groundwater available. Increasingly important is the facility to map the extent of saline and contaminated plumes and to monitor the advance of these with time.

Airborne geophysical techniques are also increasingly applied to groundwater studies. Throughout the world extensive coverage of aeromagnetic and electromagnetic data already exists. Such data, usually collected during programmes of geological mapping or mineral exploration, is often readily and cheaply available from Government departments. Until recently airborne surveys have been considered too costly to commission specially for groundwater projects but with the increasing demand for large volumes of potable water and the increasing scarcity of such supplies this attitude is changing.

1.2 Review structure

The Preface and Introductory Section, common to this series of BGS/ODA technical reports, briefly describe the scope of other reports in the series and outline the worldwide occurrence of unconsolidated aquifers.

Section 1 states the aim of the present review covering the application of surface and airborne geophysical techniques to the exploitation of unconsolidated aquifers.

Section 2 outlines both the physical principles of the main geophysical techniques and the standard field procedures and data processing and interpretation methods. This section has been included for the convenience of the reader who is not familiar with geophysics; it is necessarily brief and those requiring more detailed knowledge are advised to consult one of the standard textbooks (eg Applied Geophysics by Telford et al, 1985 or An Introduction to Geophysical Exploration by Kearey and Brooks, 1984).

Section 3 is the main body of the review and it describes how individual techniques and combinations of techniques have been applied to give some of the answers required for the successful exploitation, protection and management of unconsolidated sedimentary aquifers. This section is sub-divided according to the main issue to be solved (such as determining the thickness of unconsolidated deposits) and further subdivided corresponding to the geophysical techniques applied. This part of the review is based on published case histories that have been located following exhaustive computer-based bibliographic searches. Attention has been focussed on recent papers (1980 onwards) but reference is occasionally made to "classics" where these have not been superseded. A wide range of applications and geophysical techniques is described. I have also attempted to include studies encompassing a broad geographical and hydrogeological spread, however, by the nature of things, the majority of the case histories are USA based.

An acknowledgement is included as Section 4 while Section 5 comprises the bibliography on which the review is based.

Finally, Section 6 comprises summary sheets that show a suggested procedure for tackling the various issues to be solved by hydrogeophysical techniques. It is emphasised that these are suggested procedures; they should not be treated as inflexible and they should not be followed blindly. It is also highly desirable that geophysical procedures, from field observations through data processing to interpretation, are at least guided by (if not undertaken by) a geophysicist. These summaries are concluded with a list of key references that describe the procedures in some detail.

2. THE PHYSICAL PRINCIPLES OF RELEVANT GEOPHYSICAL TECHNIQUES

2.1 The gravity method

The object of a gravity survey is to measure horizontal variations in density. The gravity method is commonly applied at the reconnaissance stage in the exploration of large sedimentary basins, both to outline their form and to locate zones of shallow bedrock and the thickest sequences of infill. The method is also frequently applied to tracing buried channels incised into bedrock. Its undoubted success in these applications results from the large density contrast that typically exists between bedrock and unconsolidated deposits. As a rule of thumb the basins should display relief of at least 50m so that sizeable anomalies will be observed using standard gravity techniques. The method also gives best results in topographically subdued areas, remote from mountain chains, where large corrections are not required.

The earth's gravity field ranges between $9.76ms^{-2}$ and $9.83ms^{-2}$ over the surface of the globe. This range reflects variations due to elevation, latitude, local topography, the position of the moon and, most significantly, contrasting rock densities. The variations of the acceleration due to gravity, g, are measured with an extremely sensitive meter, something like a spring balance, that can resolve differences in g to 1 part in 10^8. Following numerous corrections to compensate for observable factors (elevation and latitude differences for example) the data are presented as Bouguer anomaly maps or profiles that depict anomalies reflecting changes in the density and distribution of the underlying rocks. Such changes are expressed in milligals (mGal), where 1mGal is equal to $1.0*10^{-5}ms^{-2}$.

Quantitative interpretation is usually confined to determining the distribution of rock types (and hence defining basins, incised channels etc) along characteristic profiles, normal to the strike of significant anomalies, assuming infinite strike length. 3-D interpretation is possible but, except for the simplest of models, demands large computing capacity and may not be worthwhile. It is important that realistic density values are used for the various lithologies included in the model; these may be obtained from borehole logs, from laboratory determinations or by the Nettleton (1939) field profile method.

The availability of cheap, powerful and portable personal computers has led to a revival of interest in the gravity method since the formerly onerous data reduction and modelling can now be performed largely by machine. In addition to this, as Angelillo et al (1991) have demonstrated, in groundwater surveys the correction procedures can be much simplified due to the relatively large anomalies anticipated.

Advantages of the gravity technique include one-man operation, the possibility of making observations in built-up areas and along existing roads and tracks (where levelling information already exists). The main disadvantage is the high cost of the gravimeter, typically around £40 000.

2.2 The seismic methods

The seismic refraction method has long been applied to groundwater exploration problems such as determining the depth to the water table, the depth to bedrock and even mapping the thickness of an aquiferous layer. However the technique suffers certain limitations (described below) that restrict its routine application . Seismic reflection, for long the preserve of oil explorationists, is being increasingly applied at a small scale to shallow exploration. The reflection technique offers very fine bed resolution and is not subject to the limitations of the refraction method.

2.2.1 Seismic refraction

The thickness and acoustic velocity (and hence an indication of the lithology) of individual rock layers are determined by timing the arrival of an artificial disturbance at a series of sensitive motion detectors (geophones) set on the surface. Figure 1 (a) illustrates the simple case of one horizontal interface beneath which the velocity increases. An elastic pulse is generated using a hammer and plate, falling weight, detonator or larger explosive charge etc and the radiating energy from this, travelling by several paths, is detected by the line of geophones. Some energy travels directly through the upper layer, some is reflected from the interface back to the surface while some (striking the boundary at the critical angle) is critically refracted at the interface and travels along this at the higher velocity of the deeper layer and generates head waves in the upper layer that can be observed at those geophones sufficiently distant from the shotpoint.

The time lapse between the shot firing and the first arrival of energy at each geophone is plotted against the distance of the geophones from the shot point to form a time-distance graph (see Figure 1 b). The energy arrivals at geophones closest to the shot will have travelled directly through the surface layer. However, beyond X_c, the cross-over point, the headwave generated by energy travelling in the higher velocity medium, will arrive first. The reciprocals of the gradients of the straight segments give the velocities in both the upper and lower media while the thickness of the upper layer may be determined from either the intercept time (t_o) or the critical distance (X_c). For the single layer case depth estimates are typically accurate to +/-10%.

The procedure can be extended to several layers but the interpretation becomes increasingly unstable for more than three interfaces since errors are compounded. There are also procedures to deal with irregular interfaces, where the depths to the boundaries can be calculated beneath each geophone. This can be achieved, for instance, using the plus-minus method of Hagedoorn (1959). However, these procedures require at least that the spread is shot in both directions (forward and reversed) while the more sophisticated interpretation technique (the Generalised Reciprocal Method of Palmer, 1980) demands geophone spacings that are probably too small for routine hydrogeophysical work.

There are three fundamental requirements for successful seismic refraction:

1) The layer velocities must increase downwards (preferably by a factor of 2 or more)

2) The layers should not dip at more than about 20 degrees

3) The layers must exceed a certain minimum thickness to be detectable.

Where these conditions are not met (which is quite common) the interpreted depths are subject to considerable error. A further disadvantage of the refraction technique is the long geophone spreads required relative to the depth explored; typically a spread of some 5 to 10 times this depth is necessary.

Using a sledge hammer or falling weight energy source the maximum depth of investigation is some 35m but this can be extended through the use of modern equipment that is capable of signal stacking and hence signal enhancement. Much greater depths can be explored by using explosive sources but this option has attendant logistical problems

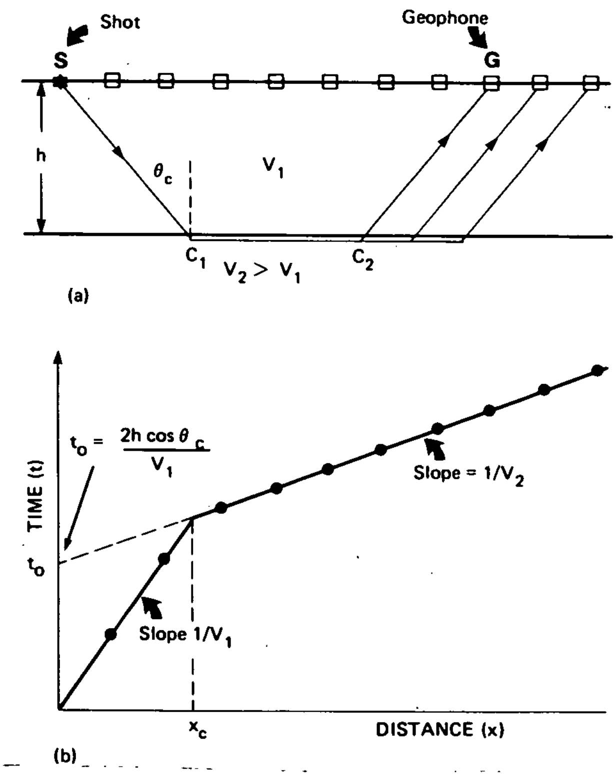

Figure 1. (a) Refraction at a single horizontal interface and (b) associated time-distance graph (from Barker, 1986)

involving permitting, storage and security etc.

2.2.2 Seismic reflection

Shallow, high resolution seismic reflection techniques have been developed during the past decade, following the introduction of digital engineering seismographs with enhancement and filtering capabilities and the ready availability of powerful microcomputers. However, these techniques are still not generally considered for routine hydrogeophysical surveys, partly through lack of experience and partly because of the (largely assumed) lengthy data processing required. However, publications by specialists in the field (the Geological Survey of Canada and the Kansas Geological Survey) have shown that the data acquisition and processing is not too onerous.

Depths in the range 3m to 200m may now be explored with standard equipment. A major advantage of the technique is its exceptionally high resolution that results from the use of high frequency sources; beds as thin as 50cm may be detected in good conditions. The other major advantage over the refraction method is that the velocity of individual beds is not required to increase with depth; seismic energy is reflected back to the surface from any boundary that displays a contrast in acoustic impedance (a parameter based on the velocity/density product of the lithologies involved). Hence the range of geological conditions where seismic reflection may be successfully employed is far greater than for the refraction technique. It is clear that shallow seismic reflection will play an increasingly important role in exploration geophysics in the future.

Two contrasting techniques are used to undertake the surveys; these are the Common Depth Point (also known as Common Mid Point and Common Reflection Point) and the Optimum Offset. The latter technique is by far the simplest (and quickest) field technique and requires less data processing to achieve a useable seismic section. These techniques are described concisely in excellent papers by Pullan and Hunter (1990), Steeples and Miller (1990) and Haeni (1985). Meekes et al (1990) give an excellent review of optimal field and processing parameters. A brief outline of the techniques is included here for convenience:

Common Depth Point (CDP)

In this method a variety of source-receiver separations (typically between 3m and 20m) are used successively so that multiple reflections are received from a single point on the reflecting horizon (see Figure 2); typically up to twelve pairs (12 fold CDP) of source-geophone separations are shot, resulting in 12 traces with information from the same point on the reflector. The entire spread is then moved one station interval (say 5m to 10m) along the traverse and the process is repeated. Kopsick and Stander (1983) give typical survey parameters for undertaking the study of an alluvial aquifer some 20m thick. The individual traces with information from the same part of the reflector are gathered together for processing, the aim being to enhance true reflections at the expense of extraneous noise. Following a correction for the different path lengths involved (normal move out) the traces are added together so that, for instance, information from the 12 traces is compressed into one trace; in this manner genuine reflections are enhanced. Then follows various low cut

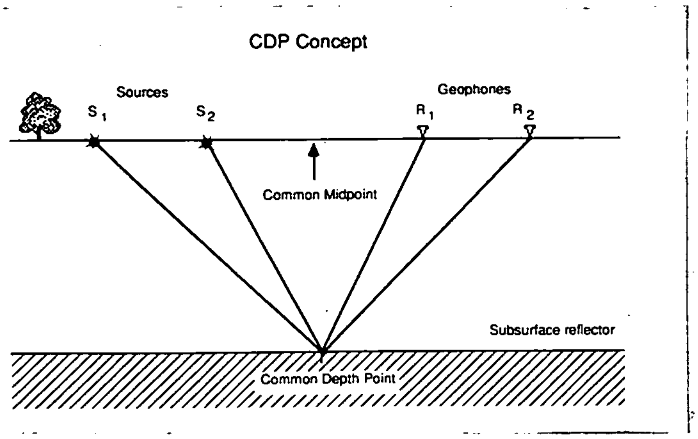

Figure 2. The seismic reflection common depth point field arrangement (from Steeples and Miller, 1990)

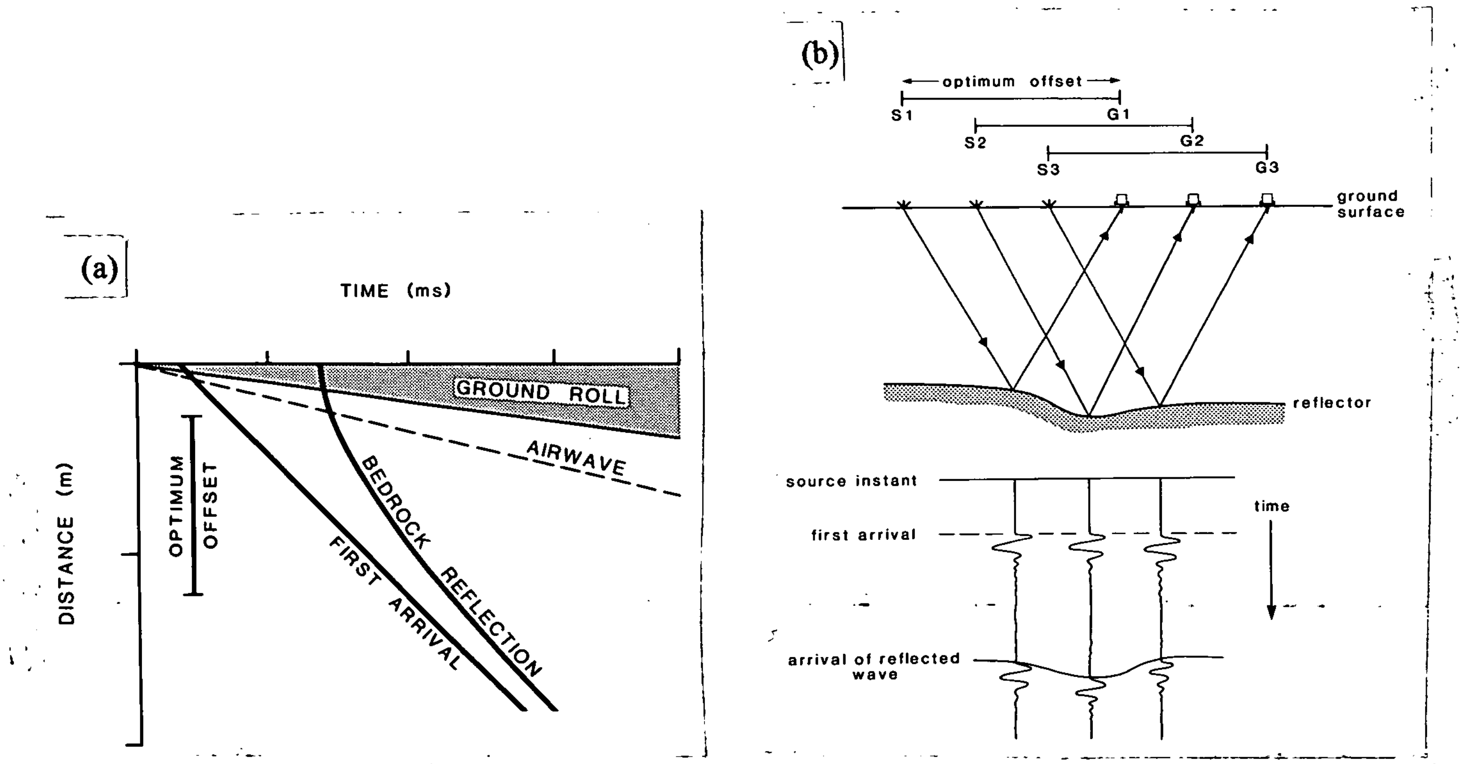

Figure 3. (a) The seismic reflection optimum offset window and (b) the optimum offset field arrangement and three schematic traces yielded by shooting from S_1 to G_1 etc. (from Pullan and Hunter, 1990)

filtering operations (to remove frequencies below about 200Hz that reflect ground roll, shot noise etc) and automatic gain control to enhance weaker reflections. Clearly the field and processing operations with the CDP technique are rather lengthy, however, Steeples and Miller (1990) state that their 4 man crew can shoot between 500m and 800m per day of 12 fold CDP coverage with shotpoints at 1m intervals. The seismic source used for this work was a hunting rifle, enclosed by an air-blast containment, and discharged into the ground.

Optimum Offset (OO)

The optimum offset refers to the surface separation between the source and the single geophone where the energy arrivals at the geophone can be clearly differentiated and the reflection from the target can therefore be isolated. This is demonstrated in the time-distance schematic of Figure 3 (a), which shows a range of optimum offsets (known as the optimum window) where the (bedrock) reflection is clear of both ground roll, airwave and direct or refracted arrivals. In a new area, the optimum offset is first determined by shooting a series of records with variable shot to geophone array distances, and then the survey proceeds by maintaining the optimum separation between the source and geophone (typically between 10m and 30m) and moving along the traverse at a constant interval (typically between 3m and 6m), as shown in Figure 3 (b). Each trace of the final seismic section is obtained by recording the output of the single geophone. Data processing is minimal and comprises static corrections (usually achieved by "hanging" each trace on the water table reflection, so that each trace is aligned on the same datum) followed by filtering etc. Finally the two-way travel times of the seismic section are converted to depths using local velocities determined by occasional seismic refraction spreads. Typical daily progress for a three man crew is 500m of seismic section with a station interval of 3m.

Common to both techniques is the use of high frequency geophones (natural frequency 100Hz) and, where possible, sources with a dominant high frequency (200+Hz); such sources include a rifle bullet shot into the ground through a muffling barrel to reduce air waves. The high frequencies employed allow severe low frequency filtering (hence eliminating traffic and power cable noise and ground roll) and permit the resolution of thin beds, the resolution limit being about 1/4 of the source wavelength. The best results are obtained in areas most favourable for the transmission of the high frequency energy, ie fine-grained and damp soil; poor results must be expected where the overburden consists of dry, coarse-grained material. It is noteworthy that the optimum field conditions for seismic reflection and ground probing radar (see below) are almost mutually exclusive.

2.3 The electrical methods

The electrical methods are particularly suitable for groundwater studies because the resistivity/conductivity of a rock largely reflect its porosity and permeability, its degree of saturation and the fluid conductivity of its pore water. Hence electrical surveys can indicate groundwater quality and content and, in certain uniform conditions, can also yield an indication of transmissivity etc. In most rocks electricity is conducted electrolytically through the interstitial fluids and hence it is the conductivity of these fluids that controls the rock resistivity rather than the resistivities of the rock matrix. The situation is complicated by the presence of clay minerals, however, because such minerals conduct electricity electronically; thus current flow through clayey lithologies is both electronic and electrolytic.

The fundamental division of the electrical methods is between direct current (dc) and electromagnetic techniques (em). Traditionally the dc or galvanic systems were preferred because the equipment is relatively simple and cheap and the method is easy to understand. However with the introduction of light-weight and sophisticated electronics the em methods are rapidly eclipsing the dc techniques. EM enjoys superior resolution and more rapid data collection since electrode contact with the ground is not required. Use of the two basic techniques is not mutually exclusive however and there are several potential applications where the contrasting methods yield complementary data. This results primarily from the fact that the dc and em techniques respond best to resistive and conductive targets respectively.

2.3.1 The direct current (dc) methods

The electrical resistivity of earth materials is measured by introducing an electric current (usually switched direct current) into the ground via two electrodes and observing the resultant potential field developed across two additional electrodes. The apparent resistivity (in ohm.metres) of the volume of ground influencing the current flow is calculated by multiplying the quotient (V/I) by a geometrical factor dependant on the separations and arrangement of the electrodes used. In the rare case of homogenous ground this value of apparent resistivity is equal to the actual or specific resistivity of the ground; more often, however, it represents a weighted average value of all the influencing lithologies.

Observations of apparent resistivity variations in the lateral sense are made in the profiling mode where, usually, the entire electrode array is moved across the survey area maintaining a constant separation between the electrodes and hence investigating a fairly uniform depth. With the more rapid gradient array, however, the current electrodes remain fixed while traversing with a pair of electrodes measuring the potential difference in the central rectangular area. Normally at least two separations are employed to allow some depth discrimination, the larger separations yielding information from greater depth. The results, plotted as profiles or iso-resistivity maps, are usually interpreted only qualitatively, to indicate contacts between units of different resistivity, fault zones (usually relatively conductive) and variations in overburden thickness.

Microprocessor controlled traversing systems are becoming popular. These require a large number of electrodes set out on a profile and linked by multi-core wire. The computer is able to address sequentially pairs of electrodes as either current (senders) or potential (receivers) as required and hence data can be rapidly acquired for a large number of arrays and separations. Pseudo-sections derived with such systems may be readily interpreted to yield both the lateral- and vertical distribution of resistivity (eg Olayinka and Barker, 1990). Rectangular grids of electrodes may also be employed; such arrangements are used to monitor resistivity variations with time across large areas which may in turn reflect the advance of contamination etc.

The vertical resistivity distribution at a site is mapped using the sounding mode (VES) which involves increasing the current penetration by progressively increasing (usually in logarithmic intervals) the current electrode separation about a fixed centre point. With the Schlumberger array, the most commonly used sounding array, the potential electrode separation is kept constant until the potential difference becomes too small to measure accurately, at which stage the separation is increased. A major advantage of this array in semi-arid areas is the reduced number of electrode moves required to complete a sounding.

Another popular array is the Wenner, in which the separation between all four electrodes is kept the same; when used in conjunction with the Barker constant offset cables (Barker, 1981) lateral effects may be isolated and a sounding may be rapidly undertaken by only one operator.

Whatever array is used the electrodes should be expanded parallel to geological strike and topographic features where possible, while fences, ditches and powerlines should be crossed at right angles. The calculated apparent resistivity values are plotted against the respective electrode separation to produce a sounding curve. This is usually interpreted in terms of layer thicknesses and resistivities by comparing the observed curve with one derived by computer for a specific model, adjusting the model until a close fit is achieved.

Two problems confronting the interpreter of VES curves are equivalence and suppression. Equivalence is the condition where a large number of quite different geoelectrical arrangements yield practically identical sounding curves. Suppression occurs where a lithological unit has either insufficient thickness (in relation to its depth) or resistivity contrast to be resolved on the sounding curve. To help resolve these problems it is important to incorporate all available data (borehole logs, observed geology etc) in an interpretation. In addition, the use of other techniques (eg induced polarisation or electromagnetic sounding) may help resolve ambiguities through constrained joint interpretations. In the complete absence of additional information the equivalence problem can be partly circumvented by restricting the interpretation to the derivation of longitudinal conductance (T/R) and transverse resistance (T*R) for each of the observed layers, where T and R are, respectively, layer thickness and resistivity. Maps showing the variation of these parameters can have great hydrogeological value and yet the interpretation to this stage requires information on neither layer thickness nor resistivity.

The dc resistivity techniques are relatively labour intensive and slow. A further disadvantage is that large electrode separations are required relative to the depth of investigation (typically 10 times this depth). Because of this, large volumes of ground are sampled and this leads to loss of resolution while rendering the technique susceptible to lateral resistivity variations.

2.3.2 Induced polarisation (IP)

This technique was developed primarily for metallic mineral exploration but since the mid 1950s it has occasionally been applied in hydrogeophysical surveys. The method relies on the build up of an electrical charge on the surface of mineral grains of metallic lustre or, more importantly for the hydrogeophysicist, on the surface of disseminated clay particles, during the passage of direct current over a period of a few seconds. When the direct current is terminated the slow dissipation of this charge back towards a neutral state is measured with porous pot electrodes on the surface. The magnitude of the IP effect is commonly expressed by a parameter known as chargeability which is defined as the area under the residual voltage decay curve normalised by the voltage across the measuring electrodes during the charging period. The IP effect can also be measured as a change of earth resistivity as measured with two different low frequency alternating currents. IP, like resistivity, can be measured in both the profiling- and the sounding modes and numerous different electrode arrays may be used.

The principal uses of IP in hydrogeophysics are to differentiate between a) dirty- and clean sandy aquifers and b) low resistivities caused by saline water and wet clays. In addition, layers that do not display resistivity/conductivity contrasts may be resolved by IP.

The equipment is rather bulky and expensive and the field procedure is slow; however, resistivity data is collected at the same time. Both profiling and sounding data can be interpreted quantitatively.

2.3.3 Electromagnetic (EM) methods

The traditional EM techniques (frequency domain EM (FEM)) employ a continuous fixed frequency signal, typically in the range 100-8 000Hz. Occasionally measurements are made at several different frequencies. The receiver coil detects the directly transmitted field combined with any secondary fields generated by conductors within the ground. The effective depth of penetration is controlled by factors such as the distance between the transmitter and receiver coils, transmitter frequency, coil orientation and ground conductivities. Interpretation is typically qualitative, outlining shallow conductivity distributions that may reflect the presence of fault zones, clayey horizons and contaminant plumes etc.

An alternative approach, based on a pulsed source (TEM), has become increasingly popular with the advent of more sophisticated instrumentation. In TEM the receiver measures the decay of a transient field in the absence of the primary signal. The magnitude and rate of decay of the transients provide information on the variation of conductivity with depth and the use of time in this context can be considered as analogous to the electrode spacing in VES.

The FEM and TEM techniques are largely complementary; FEM investigations are restricted to the top 50m or so while targets below this depth are best detected using TEM.

2.3.3.1 Frequency domain EM (FEM)

Horizontal Loop EM (HLEM)

The equipment comprises a transmitter and receiver loop linked by a reference cable and separated by between 10m and 100m. The loops are held horizontal (vertical dipole) and observations are usually made at several different frequencies. The receiver coil measures the in-phase and out-of-phase component of the transmitted vertical magnetic field. The technique was developed to locate steeply dipping conductors (mineral veins) but by varying the frequencies and coil separations, the vertical distribution of conductivity may also be mapped.

Terrain Conductivity measurements (TCM)

The EM34 ground conductivity meter made by Geonics of Canada has become widely used in groundwater studies during the past decade or so. This instrument comprises a transmitter and receiver coil in the form of moulded portable loops that are moved together, a fixed distance apart, along the traverse. The received field is measured relative to the primary field as transferred directly through a reference cable. The instrument displays the out of phase response converted to a scale of apparent conductivity and uses the in-phase response to indicate when the spacing between the transmitter and receiving coils is near the correct value of 10m, 20m or 40m. By switching the transmitter frequency automatically from 6400Hz, through 1600Hz to 400Hz according to the coil separation, the induction number is kept about constant and low enough for the linear conversion of the out of phase value to ground conductivity. Readings are made with the coils co-planar, either horizontal or vertical, giving respective penetration of approximately 1.5- or 0.7 times the coil separation. (ie ranging from c 7m to 60m). It is usual to repeat a traverse using several different coil separations and/or orientations to derive information on the conductivity distribution with depth. Such data may be crudely interpreted quantitatively to yield a series of 1-D geoelectrical sections. A sister instrument, the EM31, comprising transmitter- and receiver- coils mounted on a rigid 4m boom, allows single operator measurements of conductivity down to about 6m depth.

The Very Low Frequency (VLF) technique

The VLF method exploits signals in the frequency range 15kHz to 25kHz that are broadcast by powerful military transmitters and can be detected in most parts of the world. Either or both the magnetic field components (in phase and out of phase) and the horizontal electric field component of the remote transmissions can be measured. In the former case conductive zones are indicated by distortion of the horizontal and linearly polarised primary magnetic field. Buried-conductors cause this field to become elliptically polarised and the major axis to tilt with respect to the horizontal. The exploration depth for typical overburden resistivity (200ohm.m) is about 30m.

The VLF resistivity equipment measures the local ratio between the horizontal VLF electric and magnetic fields to derive a magnetotelluric apparent resistivity. Also measured is the electrical phase angle between these two components and this permits detection of an electrically layered earth. The data is displayed directly as apparent resistivity and phase angle.

VLF equipment is relatively cheap and the technique is rapid; a single operator can cover several kilometres in a day.

2.3.3.2 Time domain (pulse) EM (TEM)

By abruptly turning off a steady (square wave) current flowing in a large loop on the earth's surface, a transient electromagnetic field is created. This in turn induces secondary electric currents (with associated magnetic fields) to flow in horizontal circles under the transmitter loop. The decay, with time, of the vertical magnetic field component of the induced currents is measured by a smaller multi-turn loop, usually situated within the source loop. The rate of decay of the vertical magnetic field is a function of the electrical conductivity of the earth under the loops. With longer times after current switch off the induced fields have penetrated further into the earth and later measurements are therefore representative of greater depths, this relationship being governed by the distribution of conductivities present (see Figure 4).

The results are plotted in similar fashion to a standard VES, with apparent resistivity (derived from the voltages induced in the receiver coil by the secondary magnetic fields) as ordinate and time (equivalent to depth) as abscissa (cf electrode separation). Forward modelling is used to calculate the response expected over a given 1-D geo-electrical section; inversion modelling is limited to matching the observed curve to the case of a few horizontal layers of variable thickness/conductivity.

Typical transmitter loop sizes are in the range 50m by 50m to 200m by 200m, with current in the range 3A to 10A. The depth of investigation is limited by the time after switch-off that the decaying signal can be measured above noise; this can be extended by increasing the current, by increasing the loop dimension (since signal strength is proportional to the product of loop area and current flowing) and by stacking many transients. Penetration down to about 500m is routinely achieved. There is also a minimum depth of investigation with TEM, presently about 5m, due to the problems of measuring the transient field immediately following current switch off. The field operation typically requires two or three workers who can measure between 15 and 30 soundings per day.

The TEM technique is operationally superior to VES since there is no requirement for galvanic contact with the ground, nor for very large electrode spreads. Hence TEM soundings may be made in relatively confined spaces. Because TEM samples a relatively small volume of ground for any given depth of investigation, the results are less likely to be degraded by lateral insignificant conductivity variations and hence TEM offers enhanced resolution of layers compared with VES.

2.3.3.3 Airborne EM (ABEM)

These systems were originally developed for massive sulphide exploration but recent advances in both instrumentation and processing have enhanced their detection capabilities. Palacky (1989) reports that ABEM systems now respond to conductivities in the broad spectrum from 0.1mS/m to 100mS/m (ie ranging from gravels/sands to the most conductive clays). Hence ABEM surveys are now routinely applied to geological mapping,

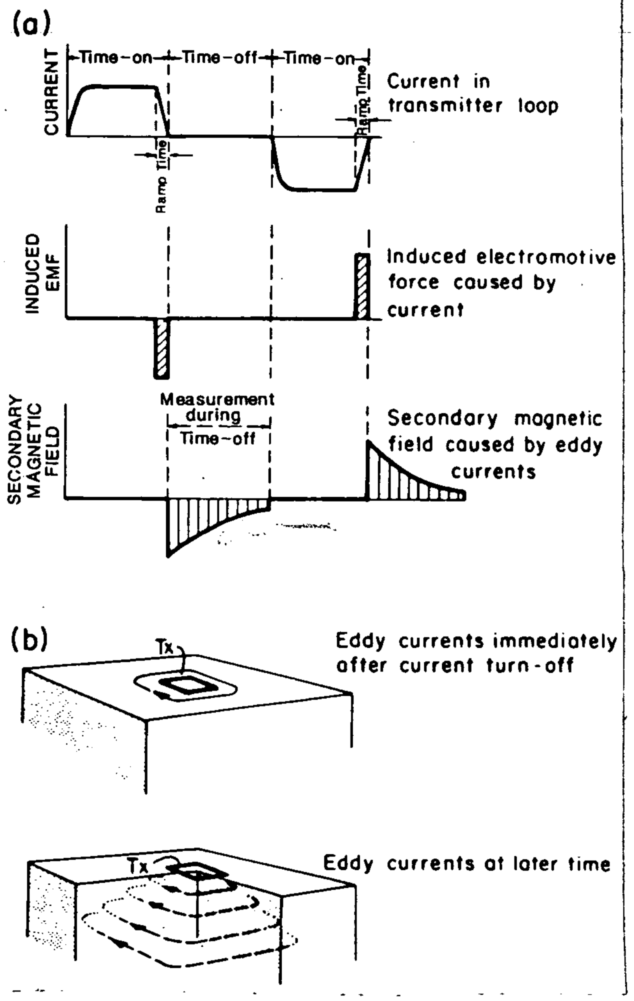

Figure 4. (a) Time domain EM waveforms and (b) homogenous halfspace eddy current flow (from McNeill, 1989)

including the definition of the extent and thickness of Quaternary deposits. Thicknesses are determined on the basis of the variable responses at contrasting frequencies. ABEM employs both FEM- (using helicopter borne multi-coils) and TEM- (towed bird) techniques. As mentioned earlier, a great deal of ABEM data has been acquired and may now be available for a nominal charge.

2.3.4 Ground Penetrating Radar (GPR)

Ground Penetrating Radar (GPR) is one of the most recent geophysical methods and was developed from experiments to find the thickness of ice sheets. In principle it is similar to the seismic reflection and echo sounding techniques. The system comprises a transmitting and receiving antennae and a PC based control console. The antennae are separated by a short distance (up to a couple of metres) and are traversed over the ground to be surveyed. The measurements may be made continuously, as the antennae are dragged across the surface, or at discrete intervals, usually of about 0.25m to 1m.The transmitter produces a short pulse (at the rate of about 50 000 per second) of very high frequency (10MHz - 1000MHz) electromagnetic energy that is directed into the ground and partly reflected across any boundary where there exists a measurable contrast in the electrical properties, principally resistivity and dielectric constant. These properties are controlled primarily by the water content and GPR is highly sensitive to very minor differences in water content and hence to very subtle lithological changes. The reflected signals are detected by the receiving antenna where they are amplified, digitised and stored on diskette ready for data processing and display. Following a simple field experiment to determine the average speed of transmission of the electromagnetic energy through the lithologies investigated, the two way reflection times are translated into depths. Alternatively this conversion may be made by calibration using borehole information.

GPR offers rapid coverage and very high resolution with detection of features a few centimetres thick at ranges of several metres. The higher the frequency employed, the better the resolution but, as a trade off, the poorer the penetration. Wide experience suggests that a frequency of 100MHz offers the best compromise between resolution and penetration. The penetration is governed primarily by the shallow conductivity conditions; in damp clays penetration is restricted to one or two metres only compared with 40m in resistive, massive rocks. Other limitations include the presence of extensive metal structures that act as extraneous reflectors and, of course, local radio frequency interference.

GPR surveys have been used to map depth to bedrock, to delineate alluvial stratigraphy in great detail (differentiating between aeolian and riverine deposits, for instance), mapping the water table depth in coarse sediments and defining the extent of groundwater contamination.

2.3.5 Electro Kinetic surveying (EKS)

EKS is the most recent hydrogeophysical technique. It offers exciting possibilities since it is the only surface technique that responds directly to the presence of moveable groundwater.

The EK phenomenon was first observed in the 1930s during seismic exploration for oil. However, it was not until 1985 that its potential for groundwater exploration was recognised (Murthy, 1985) and only in the last three years has this application been explored more fully.

The physical principles of the method, which can be considered a hybrid of the seismic- and galvanic electrical techniques, can be readily understood. Electric charge separation occurs naturally in porewater where ions (usually anions) are attracted preferentially to the surface of the mineral grains constituting the aquifer matrix. Thus the porewater exhibits a net charge and when caused to move relative to the containing rock particles an electric field is established instantaneously. Such porewater movement is induced in practice by a down-going compressional seismic pulse generated by a surface impact, usually a sledgehammer blow on a steel plate. The resulting oscillating electric field is measured across two pairs of grounded electrodes placed symmetrically about the steel plate on the earth's surface. The graphical plot of the variation of electric potential (of amplitude typically a few millivolts) across both pairs of electrodes as a function of time (measured over a total interval of about one fifth of a second from system triggering) can indicate the water table depth, the presence and thickness of saturated permeable zones, a measure of the permeability of individual horizons and hence potential groundwater yield, and the depth to impermeable bedrock. These events in time can be converted to depths by incorporating known or typical seismic velocities for the various horizons encountered. In electrically quiet conditions with good seismic coupling, depths of 100m can be examined using a hammer and plate source.

Added advantages of EKS are the lightweight and compact nature of the equipment (the complete kit can be carried in a small suitcase), its speed of operation (some 10 sites can be surveyed in one day) and its straightforward, computer-based data capture, processing and interpretation routine.

As part of the present UNSAs project, BGS has undertaken EK observations at numerous control boreholes tapping the unconsolidated sediments of the Red River Basin in northern Vietnam. We are also currently engaged in an ODA sponsored TDR project to evaluate and further refine the EKS technique for groundwater exploration in developing countries. In support of this project we have made EK observations at sand river- and collector well sites in Zimbabwe and some of our results are described in this review.

2.4 The magnetic method

Magnetometry has limited application in the hydro-geophysical exploration of unconsolidated deposits. The thickness of such deposits can be estimated where the underlying bedrock is magnetic, by interpreting the depth to the magnetic surface. There are also occasional instances of magnetite rich sands accumulating preferentially in buried valleys and such features could be traced by magnetometry.

In this technique the strength of the earth's magnetic field is measured at regular intervals using a magnetometer; local variations in this field strength reflect contrasting magnetic susceptibilities. The susceptibility of a rock, a measure of how strongly magnetised it becomes in the earth's inducing field, is determined almost entirely by its content of ferrimagnetic minerals, principally magnetite and ilmenite. Quite subtle changes in the content of these auxiliary minerals result in the very large susceptibility variations displayed by rocks and hence magnetometry, probably the most cost effective geophysical technique, can be a very sensitive tool of geological mapping. The measurements of the total magnetic field are made very rapidly using proton precession magnetometers, instruments that exploit the fact that the precession frequency of hydrogen protons is directly proportional to the ambient magnetic field. The absolute value of the magnetic field is measured, typically with a sensitivity of 1 nanoTesla (nT) which is about 1/50 000 part of the earth's field strength. The operation is very rapid with a complete reading taking about 5 seconds, and corrections to account for diurnal variations of the earth's field strength are straightforward. With some instruments there is the facility to measure the vertical gradient of the local magnetic field (across two sensors separated by about 1m); this has the advantage of both eliminating diurnal effects and enhancing shallow source anomalies.

Magnetic surveys are also conducted from aircraft, and aeromagnetic data has already been collected over a large proportion of the earth's surface. Such data reflects both lithology and structure and is therefore of importance to the hydrogeologist.

2.5 Miscellaneous techniques

2.5.1 Temperature probing

The exploration for water-bearing unconsolidated deposits using temperature profiling is described by Denne et al (1984). The temperatures are measured at regular intervals along profiles at a depth where diurnal effects are not significant, using a thermistor at the tip of a 1.5m probe. Due to the high specific heat of water, shallow aquifers will act as either a heat source or a sink and will thus appear anomalously warm in winter and cool in summer.

3 APPLICATIONS

3.1 DETERMINATION OF BEDROCK TOPOGRAPHY AND THICKNESS OF UNCONSOLIDATED DEPOSITS

3.1.1 Gravity

Gravimetric surveys have long been applied to the determination of basin shape and alluvium thickness and the complementary problem, mapping bedrock topography. Published papers describe such surveys at all scales from the truly regional, investigating major sedimentary basins, to the very local tracing of buried valleys incised into bedrock and subsequently infilled with alluvial deposits. The undoubted success of the technique in these applications results from the typically large density contrast that exists between the unconsolidated and commonly highly porous sediments and the relatively dense bedrock; such contrasts commonly exceed $0.5 g/c^3$. A further advantage is that many nationwide gravity surveys (or at least, surveys of regional extent) have already been completed and hence data suitable for reconnaissance work, already exists. Even in the absence of pre-existing coverage, the gravity technique can provide relatively cheap reconnaissance data, outlining areas worthy of detailed follow-up using other methods.

Birch (1982) describes how, as part of an hydrogeological study of the Albuquerque Basin, New Mexico, USA, gravity profiles derived from existing Bouguer anomaly maps were modelled to determine the thickness and extent of low-density Miocene to Recent basin deposits (the principal aquifer here) over an area of some 38 000km^2. The profiles were chosen to include as many existing boreholes and as much outcrop as possible, and to be largely orthogonal to the Bouguer contours. Occasional cross-lines were included for added control. Careful attention was paid to defining the regional gradient of each profile, and following the removal of this, the residual field was modelled using a classical two-dimensional method (Talwani et al, 1959), incorporating the surface topography, depth constraints from boreholes and outcrop and density values derived from local rock samples and density logs (gamma-gamma).

Following modelling of the profiles, an isopachyte map of the unconsolidated sediments and a bathymetric map of the top pre-Cambrian (bedrock) were produced. Thus the volume of unconsolidated sediments was calculated and, applying an assumed average porosity of these sediments (based on their modelled density and the grain density of quartz) the volume of available groundwater was calculated. The gravity data was also used to indicate to what extent the basin is segmented by barriers to groundwater flow and to what extent it is linked hydrogeologically to adjacent basins, both important parameters for successful mathematical modelling.

Birch (op cit) lists as the main sources of error in this gravity study: the inverse potential problem (ie numerous realistic density models will yield the same gravity profile), density variation with depth in the unconsolidated sediments (assumed constant), imperfect regional/residual separation and lack of two-dimensionality (assumed in the modelling) but nevertheless concluded that the final interpretation yielded a satisfactory approximation to reality.

Ali and Whiteley (1981) describe a similar gravity survey undertaken in the Sudan to delineate the Bara Basin and its unconsolidated fill of Tertiary to Pleistocene age overlying metamorphic and igneous Basement. Gravity observations were made at intervals of between 0.5km and 1km along six profiles approximately perpendicular to the long axis of the basin; station density averaged one per 8km^2. Regional/residual separation on the resulting Bouguer contour map, undertaken both graphically and by trend surface analysis, yielded similar results: the basin, some 45km wide, was defined by a maximum anomaly of some 30mgals. Subsequently five profiles across the basin were modelled (assuming two-dimensionality, ie infinite strike extent) using a single density contrast of $0.57 g/cm^3$. The isopachyte map derived indicates a maximum sediment thickness of 1400m and shows that the basin is probably bounded to the north and south by normal faulting but has an outlet to the southeast. Applying Gauss's Theorem (West and Summer, 1972) the authors make a simple calculation, involving residual gravity values, to derive the likely mass of water stored in the basin.

Hennon (1985) describes how iterative 3-D modelling of largely pre-existing gravity data over a $100km^2$ desert area of southwestern USA revealed two sub-basins infilled with up to 900m of partially saturated boulders, cobbles and gravel with minor sands and silt. Features of the bedrock, including its likely geological variability and topography, were also detailed.

Spangler and Libby (1968) report a gravity survey in the Basin and Range province of Arizona, USA, designed to outline zones for further investigation by seismic refraction. The alluvium filling the intermontane basin consists of very thick Quaternary and Tertiary sand, gravel, clay and caliche conglomerate that is known to be over 400m thick in places and contains large volumes of groundwater. The station density employed was $1/km^2$ and elevation and location were derived from bench marks and large-scale topographic maps. The regional gravity field was estimated by a graphical technique and then subtracted from the observed field; the resulting residual profiles indicated localised alluvium thickness in excess of 1km and also delineated bounding faults. The authors point out that gravity is a straightforward, one operator technique (in those areas where good quality topographic coverage exists) and that the cost of this non-commercial survey (covering $650km^2$) was equivalent to the cost of drilling a 150m borehole.

Tucci and Pool (1985) report the application of gravity to determine basin shape and depth in a further study of Arizona's Basin and Range Province where typically up to 3 000m of unconsolidated sediments occur with the depth to water table varying between a few metres and more than 200m. The gravity data was useful initially in a qualitative sense: the Bouguer anomaly lows outlined the deepest portions of the basins and it was assumed that the finest sediments (lowest permeabilities) would be concentrated in these zones. Quantitative modelling of depth to bedrock in these basins was hampered by both lateral density variations (reflecting the complex bedrock geology) and a vertical gradient of density within the unconsolidated sediments (reflecting compaction). Thus it was difficult to assign a suitable single density contrast at the modelling stage; nevertheless the authors state typical accuracies in determining the absolute depth to bedrock in the order of 30% and they conclude that the gravity technique, which is rapid and relatively inexpensive, provided useful information on the basin size and shape and is a valuable reconnaissance tool.

At the local scale, numerous papers report the application of gravity surveying to delineate buried valleys where coarse aquiferous sands and gravels have accumulated in these high-energy environments beneath a widespread overburden such as glacial drift or flood plain deposits etc (eg Hall and Hajnal, 1962; Lennox and Carlson, 1967; Ibrahim and Hinze, 1972 and Carmichael and Henry, 1977). In the latter paper the authors stress the importance of predictive modelling to assess the likely amplitude of anomalies for a given range of buried channel depths, depth of burial and density contrast between unconsolidated deposits and the underlying bedrock. They show, for example, that an anomaly of peak (negative) magnitude of 0.3mgal would be produced by either a channel 40m deep displaying a density contrast of $0.25g/c^3$ or by a channel only 18m deep where the density contrast is $0.45g/c^3$. They estimate that the practical limit of accuracy of a standard field survey using a LaCoste and Romberg gravimeter (sensitivity about 0.01mgal) is about 0.14mgal (requiring elevations accurate to +/- 5cm). The authors also demonstrate the importance of using analytical techniques for tidal corrections, rather than the conventional re-occupation of base stations at 3-hourly intervals.

Ibrahim and Hinze (op cit) stress the importance of removing from the observed gravity the regional component (reflecting deep and/or laterally extensive density variations) and correlating residual anomalies (assumed to reflect only variations in the depth to bedrock) with bedrock depths as proved in numerous local boreholes. In this manner it is not necessary to assume a density contrast between bedrock and overburden.

Carmichael and Henry (op cit) adopted a similar approach to interpret a gravity survey for groundwater involving 41 stations over an area of $1km^2$ in Indiana, USA where the general thickness of glacial drift is some 30m. The density of the drift was determined by a Nettleton (1939) topographic traverse. Following standard data reduction a 2-D first degree regional surface was removed mathematically. The resulting residual gravity map displayed a range of 0.3mgal and showed a sinuous low that presumably correlates with a buried valley on the bedrock surface, with glacial in-filling less dense than the adjacent bedrock. The residual gravity values were then converted to bedrock elevation by applying the Bouguer plate formula:

$$\nabla h = \nabla g / 2\Pi G \nabla p$$

where:

∇h is bedrock relief, ∇g is the residual gravity, G is the gravitational constant and ∇p is the density contrast. Using known bedrock depths from four existing wells it was possible to map the absolute bedrock depth over the entire area. Highly successful boreholes were sited at the indicated zones of deepest bedrock and interpreted depths to bedrock were shown to be correct to within half of one percent.

Culek and Palmer (1987) report the interpretation of a gravity survey to map bedrock topography in Ohio, USA, by extending across the survey area the correlation observed at 40 sites between the depth to bedrock (as proved in wells) and the coincident Bouguer anomaly. The authors point out that this interpretative technique, unlike conventional computer modelling, is independent of measured or assumed density values.

Cornwell (1985) describes the delineation of buried "tunnel valleys" in the chalk of East Anglia, infilled largely with clays. This case is unusual because the buried valleys are characterised as linear Bouguer anomaly highs, the density contrast between chalk and the infilling clays being $-0.26g/c^3$.

Angelillo et al (1991) demonstrate that so called "gravimetrie expeditive" can be usefully applied to locate palaeochannels in virgin areas. They demonstrate that many of the routine corrections of gravity surveying may be omitted entirely or modified thus improving the efficiency of the technique when applied to hydrogeology. For example, terrain corrections can be ignored where topographic maps do not exist; latitude corrections may be made relative to a local, convenient datum; the survey need not be tied to a gravity network station (ie all values determined remain relative as opposed to absolute values) and lastly, frequent returns to a base station need not be made with modern (low drift) gravimeters if tidal corrections can be taken from published tables. The authors show examples from Niger where numerous buried channels with relief up to 30m were located by rather small (-0.6mgal), but clearly defined gravity anomalies.

3.1.2 Seismic refraction

Seismic refraction is probably the second most commonly applied geophysical technique for the determination of alluvial thickness; again its success and hence its popularity derives from the typically large (positive) contrast between the seismic velocities of unconsolidated sediments and underlying bedrock. Characteristic velocities in unsaturated unconsolidated deposits lie in the range 0.3km/s to 0.6km/s, compared with 2.5km/s to 4km/s for sedimentary rocks and 5km/s to 6km/s for common crystalline rocks. Saturation of the unconsolidated deposits results in a dramatic (up to threefold) increase in their seismic velocity and hence allows the technique to detect the water table depth, as will be discussed later (Section 3.3.2).

As with the gravity method, occasional surveys of regional extent have been undertaken; these are directed primarily to investigations of the earth's crust and upper mantle but have also yielded useful information on sediment thickness. Pakiser (1976) describes such a survey to investigate the 1300km^2 Mono Basin in California. Shots of up to 400kg of dynamite were fired into geophone spreads of 2.5km from up to 25km distant. Delay time maps were plotted for both shot point and receivers from which it was deduced that the thickness of recent alluvium in the central part of the basin lay in the range 2km to 2.5km. Clearly such surveys are the exception and most refraction work has been performed at a much smaller scale, to investigate specifically the thickness and nature of sediments to a maximum depth of some 100m.

At the local scale, an early investigation of alluvium thickness (and water table depth) by seismic refraction is described by Duguid (1968). Using only a single channel seismic timer (ie an instrument recording only the time between shot and first arrival) with a hammer and steel plate source, he recorded good quality three layer graphs (dry alluvium, saturated alluvium and bedrock) that he interpreted by the intercept-time method. In some areas Duguid's interpretation showed consistently too great a depth to bedrock; in these areas the bedrock was weathered significantly and the seismic technique was detecting the depth to fresh bedrock.

Wallace (1971) describes seismic refraction surveys to determine the depth to bedrock and the water table (see Section 3.3.2) in a deep alluvial basin with principally igneous and volcanic bedrock, in Arizona, USA. Interpreted velocities of the major units present (ie dry alluvium, saturated alluvium and bedrock) increase in the ratio 1:2.3:6.8 and this appears to be a straightforward application. Problems encountered in his work include geophone cables that were not sufficiently long to record refracted arrivals from the bedrock (up to 360m deep) and the compaction of the alluvium in the deeper parts of the basin. This resulted in an increase in their seismic velocity and hence reduced detectability of the bedrock interface. Additionally it was noted that occasional caliche conglomerates within the saturated alluvium could not be distinguished by seismic refraction due to velocity overlap. Nevertheless, depth to bedrock showed very close correlation with proved depth to bedrock at four control points and Wallace (op cit) estimates the accuracy of prediction with seismic refraction in this case to be within 4%.

Seismic refraction experiments to determine the depth to bedrock as part of a project to investigate the diffusivity of alluvium in parts of the Ohio River valley were less successful (Zehner, 1973). Here the total thickness of alluvium averages some 40m, comprising a lower unit of sand and gravel (glacial outwash deposit) overlain by silt deposits (Recent floodplain deposits) averaging some 7m thick. The sand and gravel deposits are not homogeneous but contain lenses and/or channels of clay, silt, cobbles and boulders; in addition there are occasional shallow perched water tables. Bedrock is consolidated sedimentary rock (shales, sandstones, siltstones and limestone) of Palaeozoic age. Generally good records were obtained, with charges of 1kg of dynamite yielding clear first breaks along geophone spreads of up to 370m. The results of numerous end to end spreads indicated that strong velocity contrasts exist for the main lithologies/conditions encountered ie unsaturated alluvium (0.34km/s to 0.65km/s), saturated alluvium (1.53km/s to 1.9km/s) and bedrock (3.0km/s to 5.2km/s). However, depths to bedrock calculated from the seismic data (usually by the time-intercept method and occasionally by the reciprocal time method) were commonly in error by +/- 15% to 20% (and occasionally by as much as 40%) compared to the "actual" depths proved by augering. Where the interpreted depth to bedrock is too great the discrepancy is attributed to either velocity inversions within the alluvium (reflecting its inhomogeneity) or to the fact that the augered bedrock might be weathered compared to the solid bedrock determined by seismic interpretation. Hidden layers are assumed to be responsible where the seismically calculated depths to bedrock were too shallow.. Zehner (op cit) concedes that these discrepancies could be largely resolved through a combination of more detailed seismic surveys and more refined methods of interpretation; however this would clearly increase the cost of such work. He concludes that in the present survey, augering (discounting the cost of the rig) is more cost effective than seismic refraction for determining the depth to bedrock. It is also noted, however, that this would not be true where bedrock is deeper than about 60m (the limit of augering) or where detailed mapping of an irregular bedrock surface was required.

Tucci and Pool (1985) describe the results of two refraction traverses in Arizona's Basin and Range Province, shot to determine depth to bedrock, depth to water table and the presence of any discrete layered lithologies within the 1km plus thickness of unconsolidated fill. The strong velocity contrast between this fill and underlying bedrock (typically of the order of 1:2) yielded successful determinations of depth to bedrock and structure on this surface. However, the authors point out that the technique is time consuming and expensive, especially when great depth penetration is required with attendant long geophone spreads (some 10 times the penetration required) and large quantities of dynamite (30kg per shot).

Haeni (1986) provides a detailed account of the refraction method, the physical principles involved, the field operation and interpretation (including pitfalls). He also describes a groundwater modelling study undertaken by the USGS in New England where refraction seismics were used to determine the saturated thickness of stratified drift along nine strategically placed profiles. Bedrock channels were frequently shown to be strongly asymmetric while zones that were thought to have shallow bedrock were in fact shown to be underlain by significant thicknesses of saturated drift. In addition, in several swampy areas that had limited vehicle access, the seismic refraction method was the only economical means of obtaining the required data. These results, believed to be accurate to within 10% of true thickness/depth, were incorporated into a finite-difference groundwater flow model that was used to simulate, successfully, the response of the stratified-drift aquifer to imposed natural and manmade stresses.

Ayers (1988) describes refraction work to map the configuration of Palaeozoic/Mesozoic sedimentary units beneath river flood plain deposits over an area of some 230 km^2 in Nebraska, USA. He made isolated forward and reverse shots to obtain coverage of about one station/km^2. The local groundwater level was shallow (c. 1m-2m sub-surface) and where possible the seismic profiles were shot along water-filled ditches, thereby eliminating the effect of the shallow, low velocity layer and hence improving the final interpretation. An added advantage of this ditch-shooting technique is the much improved seismic coupling between the source (usually a shot gun shell) and the earth.

Velocity ranges encountered in Ayer's (op cit) work are: shallow clay (1.2km/s to 1.25km/s), saturated sands and gravels (1.6km/s to 1.75km/s), Mesozoic and Palaeozoic sedimentary rock (2.25km/s to 3.35km/s and 3.35km/s to >4.66km/s respectively). The interpreted depths to bedrock were seen to agree to within 10%-15% to the actual depths as proved in existing boreholes; in addition, the interpreted seismic velocity of the bedrock was seen to be a reliable indicator of bedrock type and age, as indicated above. The seismic results and well-log information were used to construct a contour map of the base of the alluvial aquifer; this showed several well defined channels with intervening bedrock ridges, all previously unknown, and indicated the total volume of saturated alluvium.

Haeni (1995) reports the determination of bedrock depth (in the range 14m to 75m) below stratified unconsolidated glacial drift at each of eight sites in north eastern USA and the close agreement between the interpreted depths and those proved by drilling.

To close this section mention should be made of fan shooting (Kearey and Brookes, 1984), one of the earliest and most simple refraction seismic surveying techniques. This method was originally employed to outline salt domes but was later adapted to the tracing of buried channels with their relatively slow velocity fill. A central shot is fired into a semi-circular array of geophones and the phones recording the slowest times are assumed to be those overlying the buried channel. A similar array is set up beyond these geophones, the procedure is repeated and hence the channel is traced.

3.1.3 Seismic reflection

As mentioned above, this technique is not yet applied routinely in hydrogeophysical exploration. Nevertheless, there are numerous publications proving its superior resolution of buried topography and alluvium thickness. Pullan and Hunter (1990) and Roberts et al (1992) show several examples of Optimum Offset surveys in Canada to delineate buried bedrock depressions at all scales from narrow, steep sided valleys 30m deep and only 50m, wide to broad intermontane features several kilometres wide where bedrock was clearly resolved at 300m depth. The geological situation in most of the examples is ideal, with a shallow water table and the presence of a very large acoustic impedance between the alluvium or glaciolacustrine deposits and underlying pre-Cambrian rocks; hence the high quality of the seismic sections that reveal remarkable detail of both the target interface and intra-overburden stratigraphy. Hunter et al (1989) show an example of poor reflection results where the only coherent reflection results from processing of the airwave. The main problem appears to have been the high attenuation of seismic energy in the low velocity (dry) surface sediments. A suggested remedy is to place both source and geophones in holes drilled to the water table but this would, of course, increase the survey cost.

Miller et al (1989) report a common depth point survey in Texas, USA, to map channels incised into Permian Red Beds, overlain by up to 15m of "dry" alluvium. High resolution of the Red Bed topography (to within 1m both vertically and horizontally) was required to successfully site a borehole monitoring water quality near a chemical evaporation pond. Additional work in this area revealed graben-like structures in the bedrock and further intra-alluvial features that would assist in determining the groundwater migration patterns in the unconsolidated alluvium.

3.1.4 DC electrical sounding (VES)

Traditionally the VES method has enjoyed the same level of popularity as gravity and seismic refraction for determining the thickness of alluvium. Again the success of the technique is dependant upon the large contrast in specific resistivity commonly existing between the alluvium (15ohm.m to 100ohm.m) and either well consolidated older sedimentary rocks (200+ohm.m) or massive basement lithologies (5 000+ohm.m).

A typical application of VES to determine the extent and thickness of an alluvial aquifer in the Omaruru Delta of Namibia is described by De Beer et al (1981). This survey is of further interest on two counts:

1) seismic velocity inversions resulting from occasional cemented layers within the alluvium preclude the successful application of seismic refraction. Similarly, gravity investigations proved unsuccessful because large density variations (reflecting the diverse nature of the Damaran Basement) obscured the anomalies reflecting the variation of alluvial thickness.

2) information was required on the potability of the groundwater which in turn necessitated

an electrical survey.

The delta area, some 400km^2, was covered by 170 VES (Schlumberger array) sited approximately 1km apart on eight profiles, c. 6km separate, crossing the delta area normal to the present river bed. The measuring array was expanded parallel to the present river bed (and hence presumably roughly parallel to the older channels in the bedrock). In general the area proved ideal for electrical soundings because it has minimal vegetation cover, very low relief and, locally, an homogeneous surface layer. Occasional patches of dry, wind-blown sand cover exhibiting very high resistivity (in excess of 30 000ohm.m) posed a contact resistance problem, but this was overcome by the use of multiple, long steel stakes and watering the ground around these with sea water. An initial qualitative interpretation was made by measuring graphically the total longitudinal conductance (S) at each site; the largest values of S will indicate the thickest part of the alluvium provided that the resistivity of this unit remains fairly constant. This is an entirely objective procedure and requires no knowledge of the actual geo-electrical layering present. In the present work the authors refined their interpretation by calibrating the soundings with subsequent borehole data. The resultant interpretations were used to construct a contour map of basement topography that clearly showed two deep (up to 80m) channels, neither of which correspond with the present river course. Very low resistivity values of the aquiferous layer in the northernmost of these channels indicated likely saline conditions; however, there was no indication of sea water invasion in the second channel.

Similar investigations (including local resistivities) are reported by Radstake and Chery (1992), Singh (19??), and Singh and Yadav (1982) describing respectively work in Haiti and the Gangetic alluvium of India. Parker-Gay (1972) describes a VES survey to determine the thickness of alluvial fan sediments, and hence their volume and an estimate of available water, in the high Andes of Peru. In this case the relatively resistive fan sediments were underlain by conductive (lake bottom?) sediments and, despite a certain degree of electrical inhomogeneity of the fan deposits, the resulting H-type curves were apparently interpreted successfully.

Mapping relatively resistive alluvial deposits over an area of about 100km^2 was the objective of a VES survey performed in Brazil by Rijo and others (1976). Schlumberger array soundings were made at 500m intervals along profiles crossing the area, density of coverage being about 1 sounding/km^2. Many of the resulting sounding curves were rather complex, indicating up to 7 layers, and the target horizon was frequently represented on these curves as little more than a minor inflexion. However, by incorporating parametrically defined specific resistivities and layer thicknesses proved at two control boreholes, isopachyte maps of overburden and alluvium were produced that the authors estimate to be correct to within 20%.

Alluvium thicknesses of up to 1 200m have been mapped, principally on the basis of VES survey (requiring electrode spacing of at least 6km), as part of an hydrogeological survey of the Amargosa Desert alluvial basin of southern Nevada, USA (Oatfield and Czarnecki, 1991). Important conclusions regarding water quality and aquifer transmissivity were also drawn on the basis of the VES results.

Zohdy et al (1974) report numerous successful applications of VES and resistivity traversing to outline buried stream channels infilled with coarse-grained aquiferous sediments. They note the importance of the selection of appropriate electrode separations for the traversing (based on initial VES data) and the subsequent confirmation by further VES prior to drilling.

The successful mapping of gravel aquifers in buried glacial stream channels cut into dolomitic limestone in Missouri, USA, is described by Frohlich (1974). Worthington and Griffiths (1975) successfully mapped buried pre-glacial channels cut into Sherwood Sandstone and infilled with drift deposits (clays and sands). In this case the "basement" (Sherwood Sandstone) was of intermediate resistivity .

Barker (1980) describes the rapid assessment by VES of the thickness of shallow sand and gravel deposits overlying the Chalk of Lincolnshire; the interpreted thickness was within 10% of the actual thickness at the two sites eventually drilled. Barker points out that considerable expense was avoided by undertaking the preliminary low cost geophysical investigation.

An unusual experiment in which Sclumberger array soundings were made on the surface of a tributary of the River Nile to determine the thickness of alluvium in the river bed is reported by Boulos (1972). The electrodes were anchored steel barrels, linked by well insulated cable to the measuring system in a central boat. Specific resistivities of the river water and a similar alluvium to that being investigated were determined by mini-parametric soundings to be 50ohm.m to 60ohm.m and 25ohm.m to 30ohm.m respectively. Underlying the submerged alluvium is granite of effectively infinite resistivity. In view of the experimental difficulties encountered, remarkably good sounding curves were obtained and these were interpreted to yield alluvium thicknesses that agreed to within 10% of the subsequently proved (by drilling) thickness (of the order 50m to 60m).

3.1.5 Induced polarisation (IP)

Ogilvy (1970) reports successful IP traversing in Russia to detect buried channels incised into sandstone and limestone country rock and subsequently infilled with alluvial sands containing a small proportion of disseminated clay. Conventional resistivity profiling yielded no discernable anomaly while IP anomalies up to four times background were recorded over the zones of thickest infill.

3.1.6 Electromagnetics (EM)

3.1.6.1 VLF-R

Poddar and Rathor (1983) inverted VLF-R data to yield overburden thickness. The conventional technique for such mapping in southern India was galvanic resistivity but a faster, non-contacting technique was required. The authors noted that for the case of a conductive layer overlying a resistive basement, the phase angle of VLF-R becomes sensitive to the thickness of the upper layer and by inverting this parameter (and assuming a two-layered earth) they achieved accurate mapping of bedrock depth.

3.1.6.2 TCM

Cornwell (1985) describes how in parts of East Anglia overlain by a sheet of glacial till up to 30m thick, buried valleys in the underlying chalk, infilled largely with clay deposits, were detected using 40m coil separation EM34 surveys. Using such a wide separation, the technique was able to penetrate the uniform till sheet and detect the extra thickness of conductive clays in the buried valleys.

Jones (1986) and Jones and Beeson (1988) describe a rapid method of mapping shallow bedrock depressions in Nigeria, using a conductivity apparatus (Geonics EM34). In the work reported the depressions were overlain by conductive saprolite but as the technique would work equally well in the case of an alluvial overburden, it is described here. Initial traverses were made using the vertical dipole (deep penetration) with subsequent resurveying of anomalously conductive zones using the horizontal dipole. Bedrock depressions were assumed (and, in the present work confirmed by VES) where the subsequent readings showed lower conductivities.

3.1.6.3 TEM

Fitterman (1986) reports the results of numerical modelling to assess the applicability of TEM sounding for mapping a) bedrock topography overlain by alluvium and b) porous gravels saturated with fresh water sandwiched between the alluvium and the bedrock. Fitterman (op cit) assumed resistivities of 50ohm.m, 1000ohm.m and 500ohm.m for, respectively, alluvium, porous gravels and bedrock.

Calculated curves for the simple two layer case (alluvium over bedrock) show the shape of the curves to be very sensitive to the alluvium thickness and the technique would therefore successfully map bedrock topography. The alluvium thickness can still be resolved for less resistive bedrock (and even for relatively conductive bedrock) provided that the EM transients are measured for sufficient time after current cessation (10ms in the example given). The inclusion of a gravel layer, 20m thick at a depth of 100m, had no significant effect on the sounding curve and this particular target was therefore not detectable. This work highlights the importance of predictive modelling prior to field work, thereby assuring the applicability of the chosen techniques before major expense is incurred.

Auken et al (1994) describe the mapping of water bearing glacio fluvial sand and gravel deposits in the Beder area of Denmark using in-loop TEM soundings. Using square loops of side 40m with a current of 3A they achieved penetration of up to 150m. They found that an observation density of about 16 soundings per km^2 was adequate and this number of soundings were typically completed in one day. The system responded well to the conductive surface of the Tertiary formation that marked the base of the unconsolidated sediments and the authors mapped an extensive pre-Quaternary depression that formed an important source of potable water. The nature of the infill material was also indicated by the TEM data which hence indicated zones of relatively high permeability (correlating with zones of lower conductivity).

Christensen and Sorensen (1994) report a similar application of TEM soundings in parts of Denmark to measure the thickness of glacially derived aquiferous deposits by mapping the depth to a conductive Tertiary clay forming the base of such deposits. The technique has been so successful that they are developing the concept of a vehicle towed TEM system to improve productivity (in similar fashion to the PA-CEP system for resistivity traversing).

3.1.6.4 ABEM

O'Connell and Nader (1986) report the detection of palaeochannels using ABEM while Palacky (1989) records an accuracy of +/-20% in the determination of depth to bedrock where this is overlain by Quaternary sand, gravel, till and clay in northern Ontario.

Paterson and Bosschart (1987) show data from a three frequency helicopter borne EM system that was converted to yield conductance profiles which in turn were inverted to indicate the thickness of a conductive overburden. Similarly, two frequency EM data was converted to produce a map showing conductance (the product of conductivity and thickness) of an aquifer in western USA. If the groundwater quality is constant, the higher values of conductance reflect zones of either thicker aquifer or higher porosity.

3.1.7 GPR

Davis and Annan (1989) describe GPR surveys to accurately map bedrock topography beneath up to 20m of damp sands with occasional silt and clay layers at a site in Ontario, Canada.

3.1.8 Temperature probing

Shallow linear aquiferous deposits of glacial drift in Illinois, USA, were detected by temperature profiling (Cartwright, 1968). The aquifers were characterised by negative anomalies (the largest recorded being about 2^0 C) since the surveys were made in spring when the saturated sediments acted as heat sinks. The small amplitude of these anomalies, which could easily be accounted for by other variables such as soil type etc, and the lack of information on the depth or nature of the aquifer, suggest that this technique, although of interest, will not achieve wide acceptance.

3.1.9 Integrated surveys

Numerous publications report the application of a combination of geophysical techniques; this is occasionally of an experimental nature (to determine which method is most efficient) or more usually because combined techniques are found to yield complementary information. Some of the more important integrated surveys are described briefly below:

3.1.9.1 Seismic refraction and reflection

Ayers (1989) compared seismic refraction and reflection (common depth point and common offset methods) in the study of an alluvial aquifer (Quaternary clays, silts, sands and gravels) overlying Palaeozoic- and Mesozoic sedimentary units in Nebraska, USA. The objective of the survey was to compare the efficiency of the techniques in determining the configuration of the bedrock surface, the thickness of alluvial overburden and the hydrostratigraphy of the aquifer.

In the refraction survey isolated spreads at about $1/km^2$ were shot in forward and reverse directions; additional information on shallow layers was obtained by including extra geophones at short spacings (say 2m) between the shot point and the first geophone of the main spread. A total of five short reflection profiles were shot, including at least one of the refraction stations, using both the common depth point and common offset techniques.

The refraction work indicated a straightforward 2 layer case, saturated alluvium overlying bedrock. Occasionally a surface clay deposit (low velocity) was clearly resolved. Derived depths to bedrock were generally within 10%-15% of the proved depth to bedrock and the velocity indicated the nature of the bedrock. Reflection data showed strong reflections associated with the bedrock surface and interpreted depths were consistent with those obtained with the refraction work. Further coherent reflections were observed from within the bedrock but generally there were no distinct reflecting horizons resolved in the alluvial section. Ayers (op cit) considers four explanations for this: the alluvium is geologically heterogeneous (ie no well developed, extensive layering exists), the alluvium is seismically homogenous (ie the different units display no acoustic impedance variations), the boundaries are gradational or lastly, the individual units are too thin to be resolved with the source frequency used.

In comparing the refraction and reflection techniques, Ayers (op cit) states that greater definition of the bedrock surface is provided by the reflection methods. However, as Lankston (1989) points out in his discussion of Ayer's paper, the refraction method is capable of equal resolution when subjected to a more sophisticated interpretational procedure (eg the Generalised Reciprocal Method of Palmer (1980). The refraction technique was also able to discriminate, on the basis of velocity, between the various bedrock lithologies. The main advantages of the reflection methods are that they can resolve relatively slow velocity layers underlying high velocity layers and, with a high frequency source, should be able to locate relatively thin layers. The reflection methods, however, are rather slow in operation and require fairly sophisticated software for routine field work. There certainly appeared to be no justification for applying the relatively costly reflection methods in the present survey.

3.1.9.2 VES and seismic refraction

Near Cape Flats, South Africa, the thickness of aquiferous, unconsolidated Tertiary to Recent sand deposits overlying granite and metasediments in various states of decomposition was originally investigated solely by seismic refraction (Meyer and DeBeer, 1981). Strong velocity contrasts exist between the lithologies investigated: dry alluvium (300m/s to 1 000m/s, saturated alluvium 1 550m/s to 1 800m/s and bedrock, in excess of 4 500m/s). Good records were obtained and the interpreted depth to bedrock usually matched well the proved depth. In some areas, however, there was an occasional hint of a layer of intermediate velocity between the base of the saturated alluvium and massive bedrock and in these areas the interpreted depths to bedrock were erroneous. Supplementary VES clearly distinguished a fourth, relatively conductive layer sandwiched between the saturated alluvium and massive bedrock; this reflected in situ weathered bedrock whose seismic velocity was so similar to that of saturated alluvium that it was generally not detected by seismic refraction.

3.1.9.3 Gravity and VES

Stanley (1972) describes a survey including gravity and VES to investigate the hydrogeological conditions in a $100km^2$ area of Cache Valley, Utah and Idaho, USA. Gravity modelling indicated the major structures and yielded thickness estimates of the saturated unconsolidated fill. A transverse resistance map for the uppermost 100m was prepared from the VES data and outlined the distribution of thick clays and sands while the presence of more resistive material at depth represented targets for deeper water wells.

Buried tunnel-valleys (narrow, steep sided channels incised into chalk and infilled entirely with boulder clay or with clay overlying sands and gravels) have been traced in East Anglia by a series of gravity profiles made normal to the present day drainage. The tunnel valley is manifested as a well defined positive Bouguer anomaly, reflecting the positive density contrast between chalk ($1.9g/cm^3$) and Boulder Clay ($2.1g/cm^3$) (Barker and Harker, 1984). Three of the profiles (one including a borehole drilled to chalk bedrock) were interpreted assuming a variety of density contrasts, the correct contrast being that for which the interpreted depth to chalk matched that observed in the borehole. A steady "down valley" reduction in computed contrasts was noted and assumed to reflect an increasing sand/gravel component of the valley fill in this direction. Thus, the authors concede, although the lateral boundaries of the buried valley are accurately delineated, the interpreted depths are likely to be considerably in error, particularly where an appreciable fraction of the infill is sand and gravel. Several short VES traverses were made subsequently and the resulting geo-electrical sections confirmed the location of the tunnel valleys as shown by the gravity surveying. However, the resistivity interpretation of the depth to chalk proved superior, resulting from the resistivity contrasts existing between chalk (50ohm.m to 70ohm.m), boulder clay (20ohm.m to 35ohm.m) and sands/gravels (90ohm.m to 250ohm.m).

3.1.9.4 VES, TEM and FEM

An example of the complementary use of VES and FEM to resolve an equivalence problem is provided by van Kuijk et al (1985) who describe VES surveys to locate borehole sites in shallow alluvial pockets overlying the Basement Complex in the Sudan. They give an example of two equivalent four layer solutions to a Schlumberger sounding, setting bedrock at either 13m or 33m deep. The expected FEM (Geonics EM34-3) responses for these two models were then calculated and compared with the actual values observed during field traverses with this instrument. This suggested that the shallow bedrock model was correct and this was confirmed by subsequent drilling.

Fitterman et al (1988) examined the equivalences displayed by three different sounding techniques VES, TEM and FEM; they concluded that since the methods are based upon different physical principles they display different but partly complementary advantages and limitations and hence their combined application is frequently indicated. In particular, for the geoelectrical arrangements that we are usually faced with in alluvial aquifers (ie H-type (middle layer conductive) or Q-type (successively more conductive with depth) they noted the following important points. VES results exhibit S (thickness/resistivity remaining constant) equivalence over H-type layering and show poor definition of the second layer (suppression) in Q-type layering. TEM is better able to resolve the S equivalence, providing more information about layer resistivity and thickness of the intermediate layer than either the FEM or VES methods. TEM has a slight advantage over the other two methods for resolving the second layer in the Q cases studied. The authors concluded that, while the combination of techniques may be successful in resolving equivalence/suppression problems, the best solution is achieved by the inclusion of additional information (from boreholes etc) or by the application of an additional (non electrical) geophysical technique.

3.1.9.5 Gravity and TCM

Further work on tracing buried tunnel valleys cut into the chalk of East Anglia showed that the combination of rapid EM conductivity mapping using the Geonics EM34 (with 40m coil separation, horizontal dipole and nominal penetration of 30m) and gravity traverses to be most effective (Cornwell and Carruthers, 1986). The valleys were clearly delineated by elongate conductivity highs, the greatest values of which generally coincided with maximum residual gravity values, ie the deepest point of the valley. The use of various coil separations enabled estimates of till thickness and composition to be made, while the EM results also helped resolve ambiguities in the gravity interpretation. The gravity traverses were well able to determine the cross-sectional form of the buried valleys. The seismic refraction technique had failed in this application due to the overlap in velocity displayed by shallow, weathered chalk and till.

3.1.9.6 Gravity and tellurics/magnetotellurics

Alvarez (1991) stresses the importance of defining the macroscopic characteristics of an aquifer prior to its exploitation; when possible constraints and limitations of the aquifer are discovered early there is less possibility of over-exploitation with its consequent often disastrous effects. Alvarez (op cit) describes a combined telluric/magnetotelluric/gravity study of 1000km^2 of the Guaymas Valley unconsolidated aquifer system of NW Mexico. This comprises essentially a two layer aquifer (sands and gravels) separated by a thick (160m) clay horizon, the whole overlying granitic basement. The telluric measurements were made at two frequencies; they yielded profiles of either relative resistivity (tellurics) or apparent resistivity (magnetotellurics) to large depths (c 2 000m) without the need for the very large surface arrays and cables that would be required for conventional galvanic measurements. These profiles were subsequently modelled to show resistivity distributions that indicated horst/graben structures in the resistive bedrock, the vertical distribution of the moderately resistive aquifers, the intervening clay horizon and additional low resistivity zones thought to reflect the presence of either hydrothermal fluids (associated with a basement uplift) or invading sea water. The gravity survey comprised randomly distributed stations (1/2.5km^2) and four profiles corresponding to the electromagnetic profiles. The resulting residual anomaly map outlined areas of shallow and deep basement and the presence of linear basaltic fissure eruptions, while modelling of the profiles largely confirmed the geological structure as indicated by the telluric/magnetotelluric modelling over this extensive area.

3.1.9.7 Gravity, seismic refraction and reflection

The source of relatively cold water feeding the Cedar Bog, Ohio, USA has been the subject of numerous geophysical investigations (Wolfe and Richard, 1990). Gravity surveying at a station density of 10/km^2 revealed a major buried valley and tributary system incised into Palaeozoic bedrock and subsequently infilled with lake sediments (silt) and glacial till. Several seismic refraction spreads were shot, primarily to determine accurately the depth to bedrock and thus constrain the 2-D gravity interpretation of selected profiles. Finally a seismic reflection profile was run; this confirmed the general shape and dimensions of the buried topography and delineated the fine glacial stratigraphy of the infilling material.

3.1.9.8 Gravity and seismic refraction

Van Overmeeren (1975) reports the combined use of gravity and seismic refraction to determine the structure and depth to bedrock in an alluvium filled basin in Chile. The gravity data were collected on profiles about 6km separate at a station interval of about 500m. Two coincident seismic refraction profiles were shot for comparison with the gravity data; modelled velocities were adjusted to provide an exact fit with depth to bedrock as proved by a single borehole. The gravity interpretation was then adjusted to match the seismic models. Two boreholes sited on the resulting geophysical indications encountered high transmissivities and much improved yields.

Van Overmeeren (1980) undertook a similar combined seismic refraction and subsequent gravity survey to delineate the course of a buried graben structure infilled with sand and gravel deposits intercalated with cellular lavas, volcanic ashes and tuffs in northern Chile. The problem at hand was not ideally suited to either technique applied: the widespread occurrence of a shallow high velocity ignimbritic sheet limited application of seismic refraction while the highly varied geology (ie large variations of densities) and rugged nearby relief (the Andes mountains) detracted from the gravity technique. However, the graben was detected following complex interpretation of one of the seismic traverses and could subsequently be traced as a distinct negative gravity feature on this and adjacent profiles at about 1km interval.

3.1.9.9 TCM and TEM

The combined application of conductivity profiling (Geonics EM34) and time domain em sounding (Geonics EM47) to map buried palaeochannel geometry in the Abu Dhabi Emirate is described by Fitterman et al (1991). Alluvial deposits comprising gravel and sand with a variable matrix of silt and clay overlie an irregular erosional surface cut into Tertiary bedrock consisting of mudstones, shales and limestones. The thicker, gravel-rich zones (infilling palaeochannels) form an important supply for nearby towns and are an important recharge zone for a shallow Tertiary aquifer. Previous VES work in this area indicated that extreme contact resistance posed a major problem, hence EM methods were chosen for this study. Initial TCM profiles were made, employing the three standard coil separations and both horizontal and vertical dipole coil orientations. Qualitative interpretation of the profiles indicated that the resistivity decreased with depth (with increasing moisture content) while quantitative, 1-D inversion at selected stations showed a three layer situation (surface 100ohm.m to 530ohm.m, intermediate 20ohm.m to 160ohm.m and bottom 5ohm.m to 30ohm.m) and outlined depressions in the lower, conductive layer (Tertiary mudrocks). Almost without exception, the bedrock channels and the present surface drainage expressions are not coincident. Quantitative interpretation of subsequent TEM soundings (at 40m interval) gave comparable indications of bedrock undulations and, by incorporating well information, enabled a basal gravel deposit (the important aquifer and recharge zone) that was originally not clearly resolved, to be delineated. The TCM data gave reasonable estimates of bedrock depths in some zones but were generally inferior to the TEM interpretation, particularly when this incorporated borehole results.

3.1.9.10 Miscellaneous integrated surveys

An integrated geophysical investigation of an unconsolidated coastal plain aquifer system and the underlying bedrock geology of New Jersey, USA, is reported by Sandberg and Hall (1990). The unconsolidated Cretaceous coastal plain sediments dip towards the southeast and range in thickness from 30m to 120m; they comprise an upper and lower sandy aquiferous layer divided by a clay-rich confining layer. Bedrock comprises pre-Cambrian crystalline and Palaeozoic sedimentary units. Seismic refraction spreads were interpreted using a ray-tracing programme that yielded a depth to bedrock under each geophone; the results of numerous spreads were incorporated with borehole results to produce a map of bedrock topography of the 200km^2 area. Bedrock lithology was mapped on the basis of long magnetic and gravity profiles, incorporating physical properties derived from borehole logs, seismic velocities etc. The contact between the upper aquifer (relatively resistive, low chargeability) and the underlying confining layer (relatively conductive, high chargeability) was accurately defined by simultaneous inversion of VES/IP Schlumberger array soundings. Elsewhere in this survey IP data helped differentiate between clay/silt layers and sands containing saline water, and also to resolve the bedrock surface beneath a conductive layer that would not have been possible by VES alone. The electrical soundings did not penetrate to the lower aquifer but, knowing the attitude of this, it was possible to predict (on the basis of the bedrock topography map) where this unit pinched out against rising bedrock and this prediction was subsequently confirmed by drilling.

Both a broad (5km) and narrow (150m) buried glacial valley, up to 60m deep, in Kansas, USA, were investigated using springtime Landsat imagery, seismic refraction, gravity, VES and temperature profiling (Denne at al, 1984). The aquiferous infill comprises Quaternary drift, loess and alluvium and bedrock is variously sandstone, shale and limestone of Palaeozoic age. Distinct tonal patterns identified on computer-enhanced imagery (particularly that taken in springtime when new vegetation growth commences) were thought to be associated with buried valleys; hence the study areas were chosen following an examination of Landsat imagery. The location of the larger valley was fairly well defined by a negative temperature (2°C) anomaly. The cross-valley profile was determined from VES and seismic refraction to within 25% of that proved by drilling. This rather disappointing correlation is ascribed by the authors to the combination of heterogenous fill and lack of a strong physical property contrast between bedrock and fill. Gravity traversing of the smaller valley yielded a very accurate cross section. The authors include an interesting productivity comparison: to drill a 90m hole requires 3man.days while a seismic profile requires 0.5man.days, VES investigation 0.5man.days and gravity data collection 4man.days, followed by some 20man.days of processing and interpretation.

3.2 DETERMINATION OF STRATIGRAPHY AND COMPOSITION OF UNSAs

3.2.1 Seismic reflection

Undoubtedly the finest resolution of alluvial stratigraphy down to depths of 100m or more, is provided by high resolution seismic reflection surveys. Pullan and Hunter (1990) show examples where optimum offset work resolved 1m thick layers of till and small lenticular pockets of sand at about 40m depth. The distinction between massive and varved clays and fine/coarse sands and gravel was also made. Clearly the technique is not able to determine the lithologies independently, but the layers are well defined and, once identified in a borehole, they may be easily traced.

A further optimum offset survey is reported by Slaine et al (1990). They were required to test, with great accuracy, the continuity of impermeable horizons in a 40m thick glaciolacustrine sequence (sands, silts and clays) overlying Palaeozoic dolerite at a site proposed for hazardous waste disposal in Canada. They used a 12 gauge in-hole shotgun source and with the dominant reflected frequencies at about 500Hz achieved vertical resolution in the range 1m to 2m. All the main lithological boundaries below about 10m were clearly defined and the two impermeable horizons were shown to be continuous across the area of interest. Comparison with numerous boreholes showed that their calculated depths to reflectors were accurate to within 4%.

Geissler (1989) describes a single-operator reflection survey to test the continuity of unconsolidated Tertiary aquifers draped over complex bedrock structures and possibly interrupted in places by steep faulting. This worked formed part of a groundwater recharge scheme whereby surface water was to be diverted to infiltrate the outcropping aquifers several kilometres from the well field where the water was to be abstracted. 4-fold common depth point data were collected and the resulting interpretation shows numerous marker horizons within the aquifers yielding strong reflections; the seismic section shows that the aquifers, down to depths of 300m, are indeed continuous. Data, acquired using 0.5kg of explosives buried some 3m to 5m deep, was generally of good quality except where the geophones had to be planted in dry sands.

Meekes et al (1990) show examples from near Utrecht, Holland, where 125m deep clay horizons in Quaternary alluvial deposits are clearly not continuous, with important implications for hydrogeological studies. Steeples and Miller (1990) present numerous examples of the resolution of intra-alluvium stratigraphy; in one instance they estimate the vertical resolution to be slightly less than 0.25m! Roberts et al (1992) give examples of optimum offset surveys that clearly define the foreset and topset beds and the distributary channel facies in the prograding River Fraser delta in British Columbia, Canada.

Brabham and McDonald (1992) detail a shallow common midpoint high resolution reflection survey made in the inter-tidal area of Barry Old Harbour in Wales. They traversed a buried valley incised in Carboniferous Limestone and subsequently covered by up to 20m of unconsolidated material (recent sands and estuarine gravels, sands and silt). They compare results achieved with both hammer and plate- and detonator sources. The detonator source yielded energy rich in frequencies between 400Hz and 600Hz, compared with a dominant frequency of about 120Hz for the hammer and plate. The much improved resolution of the fine layering within the alluvium yielded by the detonator source is striking. The authors also show that a conventional seismic refraction spread interpreted using the Generalised Reciprocal Method (Palmer, op cit) yielded the same bedrock topography as the reflection technique but did not, of course, yield any information on the intra-alluvial layering.

3.2.2 Seismic refraction

Cornwell 1985) reports the general failure of seismic refraction applied to the investigation of sand and gravel resources overlying London Clay in East Anglia: the method usually failed to accurately determine the depth to London Clay due to the presence of high velocity clay horizons within the intermediate velocity sands and gravels.

Tucci and Pool (1985) report the success of refraction traverses in Arizona in detecting coherent velocity variations in great thicknesses of basin infill (up to 1 km) that probably reflect different lithological units. The method applied at such a scale, however, is prohibitively expensive.

3.2.3 VES

Worthington (1972) achieved good results using VES to map glacial drift deposits in the Fylde region of Lancashire, UK, where he was able to resolve layers of boulder clay, sands and gravels. The shallow sand and gravel deposits are important as the loci of recharge into the underlying Sherwood Sandstone aquifer.

Geirnaert (1974) reports the successful use of Schlumberger array VES to map the depth to impermeable Pliocene sediments (in the depth range 10m to 50m) that formed the base of an important Quaternary aquifer in a deltaic region of Spain. Successful boreholes were sited on the basis of isopach and formation resistivity maps of the Quaternary deposits; zones of formation resistivity less than 30ohm.m were avoided as likely reflecting sea water contamination or increased clay content.

Worthington (1977) describes a Schlumberger sounding survey to determine the groundwater potential of the Kalahari Deposits in part of Namibia/SW Africa. These unconsolidated sediments can achieve great thicknesses (600m) and electrode separations of up to 7km were required to resolve the Damaran basement. The results indicated a quadripartite subdivision of the sediments and of these the Middle Kalahari unit (calcareous sands and sandstones) are known to constitute the major aquifer. Despite the dearth of control boreholes, it was possible to map the distribution of transverse resistance (T*R) of this Middle Kalahari unit and hence indicate the most promising drilling sites (ie in the zones of highest T*R).

Van Zijl et al (1981) report the application of high density VES (3 sites per km^2) and limited supporting techniques to the detection of alluvial aquifers (comprised mainly of sand) and their compositional variation in part of the Bree Valley of South Africa. Five major deposits, up to 40m thick, were located; not only are these important aquifers in their own right, they are also zones of recharge. Permeability variations were highlighted within the aquifers, permeability being greatest in the upper reaches and reduced (by argillaceous content) at lower levels. A well developed overlying boulder layer was detected at two of the aquifers.

Tucci and Pool (1985) report the success of large separation VES (5km) to outline zones of fine-grained sediment (characterised by specific resistivities of less than 10ohm.m) and determine their thickness in deep basins in Arizona's Basin and Range Province.

Al-Ruwaih and Ali (1986) report the use of VES to map potable aquifers in unconsolidated sediments of Pleistocene age in Kuwait, but demonstrate the fact, first highlighted by Flathe (1955), that VES are unable to resolve layers (whatever the resistivity contrast with enclosing layers) that are less thick than one tenth of their depth of burial. Thus relatively thin clay and silty layers that are of great hydrogeological significance since they divide the basin into a multi-aquifer system, could not be resolved on VES curves.

Schlumberger array VES, at approximately $1/km^2$, were used to map the extent and thickness of a shallow clay layer overlying an important alluvial aquifer system in the lower Platte Valley flood plain, Nebraska, USA (Ayers, 1989b). The presence of the clay layer could be confirmed by even a qualitative appraisal of the sounding curves; quantitative interpretation (including calibration with several borehole results) showed the resistivity of the clay horizon to be about 11ohm.m, compared with average values of 35ohm.m and 85ohm.m for the variously saturated, thin surface layer and the main sand/gravel aquifer respectively. The clay unit was present only on the west side of the present drainage, where it is areally extensive and lenticular in form, its thickness ranging between 4m and 12m, underlying up to 2m of soils. Following the accurate definition of the impervious clay layer it was possible to modify the hydraulic model (3-D, finite difference) of the entire aquifer system and hence determine the considerable effect of the confining clay on the groundwater flow system.

3.2.4 Induced polarisation (IP)

Vacquier et al (1957) experimented with the application of IP to groundwater studies in both the laboratory and field. They noted that an IP response is generated when sands and gravels in aquifers are coated with a film of clay (so called membrane polarisation); clean sands and gravels or clay slurries did not yield a response. Furthermore the decay rate of the induced potential appeared to be directly related to the average grain size of the coated grains. They also noted instances where apparent resistivity profiles remained featureless (over certain buried valleys) but a recognisable IP anomaly was generated. Thus IP appeared to have great potential for groundwater surveying. Its subsequent generally poor take-up probably results largely from the cost of the equipment and the extra time required to make IP measurements.

Bodmer et al (1968) made experimental dipole-dipole traverses of shallow (less than 10m subsurface) unconsolidated aquiferous sediments in California using frequency domain IP equipment. They noted that solid clay horizons did not yield an IP response; the largest anomalies were recorded over sands and gravels contaminated with between 3% and 20% clay particles. Thus IP anomalies indicated zones of reduced permeability. The authors point out that IP surveys are frequently able to resolve lithological contrasts that are not detected by concurrent resistivity surveys.

3.2.5 TCM

The use of EM31 and EM34 conductivity meters as sensitive mapping tools in unconsolidated overburden is described by Zalasiewicz et al (1985). The techniques yielded ready differentiation of deposits mapped as "alluvium" into sands/gravels and impermeable clays. In one instance the discrepancy between EM31 indications of clay horizons and the mapped geology led to a revision of the geological map! In addition, conductivity contour maps produced for parts of East Anglia clearly indicate the distribution of boulder clay thickness.

Potts (1990) describes the now routine application of the EM34 unit to delineate shallow "shoe-string" sand aquifers within, and overlain by, predominantly fine-grained unconsolidated alluvial bodies in Australia. This technique has now replaced dipole-dipole DC resistivity methods that were formally employed here. Coverage with the EM system is quoted as 7.5km/day, at least four times quicker than the resistivity method, and good resolution of the sandy aquifers in the top 15m to 20m is obtained.

3.2.6 TEM

Fitterman et al (1991) show how TEM soundings, initially interpreted as three layer cases, were subsequently, following the incorporation of limited borehole information, seen to represent five-layer cases. Using this refined interpretation, an important gravel aquifer (previously unresolved) infilling palaeochannels in Tertiary mudrocks, was mapped accurately.

Taylor et al (1991) report a TEM survey, comprising soundings at 75m interval on three parallel traverses, to locate fractures in volcanic flows under alluvial cover and to determine zones of high clay content within the alluvium where boreholes would probably have limited success due to low hydraulic conductivity. Using Archie's law they calculated the range of formation resistivity of productive aquifers to be 80ohm.m to 1300ohm.m, based on a range of porosity likely to be encountered in well sorted alluvium (20% to 35%), and a range of conductivity of potable water (15uS/mm to 70uS/mm). The authors accepted that this was an imperfect criterion since they had not accounted for the precise nature of the electrical conduction by various clay minerals. Nevertheless, the results from a limited amount of test drilling incorporated with pre-existing borehole information, did indicate increased hydraulic conductivity where TEM modelling showed the aquifer resistivity to be in the required range. TEM soundings on profiles also clearly outlined thick alluvial accumulations.

3.2.7 HLEM

Buried stream channels, comprising up to 20m thicknesses of coarse sands and gravels located in estuarine clays near the mouth of the River Amazon in Brazil, form relatively resistive targets that were readily located by traversing with dual frequency conventional HLEM equipment (Verma and Bischoff, 1989). The authors contrast the responses given by various coil orientations and frequencies applied, both in laboratory (physical modelling) and field experiments.

Balmer et al (1991) successfully delineated gravel-rich deposits in alluvium filled palaeochannels underlying the present drainage system in Niger using a horizontal loop electromagnetic system (2048Hz and 20m separation). The highest yielding boreholes were sited in zones displaying apparent resistivity values in the range 80ohm.m to 200ohm.m, compared with values generally less than 40ohm.m over silt/clay rich deposits. By correlation with numerous boreholes the authors were able to relate observed resistivity with permeability and hence produce a recharge map.

3.2.8 VLF

Muller (1992) reports the use of multi-frequency VLF-R measurements to map permeability variations in 25m thick Pleistocene/Holocene sands and gravels at an experimental tracer site in the Rhine valley of southwest Germany. Tracer experiments had indicated very tortuous paths through these deposits. The three VLF frequencies allowed mapping of the resistivity distribution at three discrete levels. The resulting apparent resistivity maps revealed a complex environment with the highest resistivities (assumed to reflect the highest permeabilities) following palaeo-channels, variously interrupted.

3.2.9 GPR

Very fine resolution of layers within the top 20m of an alluvial sequence in Ontario, Canada, was provided by GPR (Davis and Annan, 1989). Following careful processing of the original data, reflections from thin silt and clay bands in a generally fine sand sequence were observed and the disposition of these layers (of aeolian and fluvial origin) over a 400m traverse length is clearly defined in the radar section.

3.2.10 Microgravity

Poeter (1990) describes an unusual application of microgravity surveying before and after pumping to delineate textural heterogeneities in unconfined aquifers. The technique involves observing the differences in gravity profiles across the pumped borehole site. The largest differences indicate large drawdown (hence more permeable and higher specific yield materials). The anomalies to be expected are extremely small (calculated as only 23ugals at the centre of the cone of deposition for some 2m of drawdown) and hence great care must be exercised during the survey.

3.2.11 Integrated surveys

TCM and VES

Hazell et al (1988) describe combined TCM and VES to detect shallow aquifers in broad flats (up to 2km wide) of riverine alluvium overlying bedrock in Nigeria. The favoured targets are coarse sands and gravels of channel lag and longitudinal and point bars; these occur as lenses or nearly horizontal layers in the top 15m of sediments and are commonly obscured by overbank clays and silts. A strong resistivity contrast exists between the silts and clays (5ohm.m to 30 ohm.m) and the coarse river sands (up to 300ohm.m). Potential sites were located by rapid (1km/hr average) TCM traverses (sited using air photographs) employing Geonics EM31 and EM34. Confirmatory VES were made at these sites and the joint interpretation of TCM and VES data helped resolve equivalence problems. The authors also indicate how a qualitative assessment of aquifer thickness can be made on the basis of the ratio of the EM responses for various coil separations.

The interior structure and thickness of an extensive moraine that constitutes a major aquifer complex in Ontario, Canada, has been investigated in detail using a combination of borehole- and numerous surface geophysical techniques (Pullan et al, 1994). Shallow seismic reflection data was dominated by the response to a high velocity till (an aquitard) that was clearly defined. However, because this till reflected energy so strongly, the reflection technique could not consistently map the bedrock interface. GPR yielded fine details of shallow stratigraphy, especially where the surface sediments were resistive (sandy). TCM data (acquired using the EM34 system) provided information on the variability of shallow materials near infiltration test sites and could hence be used to assess the regional applicability of the test results in terms of recharge calculations. TEM soundings successfully mapped the bedrock (a relatively conductive shale sequence) at depths down to 150m.

TEM and VES

Christensen and Sorensen (1994) comment that TEM soundings in Denmark's important fluvio-glacial aquifers, while successful at delineating the total thickness, were generally unable to resolve the relatively high resistivity sand and gravel horizons. They demonstrated the enhanced interpretation of the internal structure of these aquifers yielded by a joint inversion of TEM and VES.

TEM and dipole-dipole resistivity profiling

Petersen et al (1989) describe combined central loop TEM soundings and dipole-dipole resistivity profiling to identify zones of well sorted coarse material down to about 300m in heterogenous unconsolidated deposits in the Basin and Range Province of Nevada, USA. They found the two techniques to be complementary: TEM soundings resolved the vertical extent of the more resistive (aquiferous) sediments while the dipole-dipole data indicated their lateral extent.

3.3 DETERMINATION OF DEPTH TO THE WATER TABLE

The electrical resistivity and seismic velocity of dry- and saturated sediments are strongly contrasting and hence the water table should generally be readily detectable by either of these techniques. This is usually so in the case of coarse sediments but in finer deposits the water table may extend across a capillary fringe that may be several metres thick. The degree of saturation throughout this fringe will vary continuously from dry to saturated and such a target may not be readily resolved by geophysics.

3.3.1 VES

Tucci and Pool (1985) report that the VES technique was unable to detect the water table in some very deep studies (requiring electrode spreads of up to 5km) in Arizona. They ascribe this failure to the heterogeneous nature of the basin fill and the general loss of resolution of the VES technique when large electrode separations are required.

Al-Ruwaih and Ali (1986) describe how VES successfully mapped the depth to the water table at about 22m in Pleistocene sediments in Kuwait; unsaturated and saturated sediments were characterised by resistivity values in the range 100ohm.m to 160ohm.m and 39ohm.m to 49ohm.m respectively. It is not clear from this paper whether their geophysical indications were tested by drilling

Haeni (1995) reports the satisfactory detection of the water table by VES in stratified glacial deposits of north eastern USA at the majority of sites tested; the technique failed where there was overlap in the specific resistivities of unsaturated and saturated deposits.

3.3.2 Seismic refraction

Duguid (1968) reports the determination of a shallow (3m to 5m) water table in alluvium accurate to within 10%, using a rather primitive single channel seismic timer. A large velocity contrast was observed between dry- (310m/s) and saturated (1 340m/s) alluvium.

Wallace (1971) reports an accuracy of interpreting the depth to the water table (up to 80m deep) in thick alluvial deposits in Arizona, USA, to be within 10%. He describes, however, instances where the seismic refraction method is not successful because the water table occurs close to the bedrock surface and/or the bedrock does not display sufficient velocity contrast with saturated alluvium. It is suggested that both cases could be investigated by modelling to determine the limits of detectability prior to field work.

Tucci and Pool (1985) report the successful application of very long spread refraction seismics to determine the water table depth at about 100m sub surface in Arizona. They report a strong velocity contrast between unsaturated/saturated basin fill (in the range 1:1.5 to 1:2).

Haeni (1986) reports refraction studies in New England, USA, where the depth to the water table (some 10m) in unconsolidated drift and alluvial deposits was successfully determined provided that the saturated section was thicker than the unsaturated. The velocity ratios between unsaturated and saturated materials were large, ranging between 2 and 6, and hence the success of the technique in this application.

Sternberg et al (1990) show examples where the refraction technique failed to determine a rather deep (180m) water table in alluvial sediments of south west Arizona, USA. They attribute the failure to the combination of the great depth of the water table and the induration of deep but unsaturated alluvium so that little velocity contrast existed across the deep water table.

Haeni (1995) subsequently reported the successful determination by refraction seismics of the shallow (1m-4m) water table depth at each of eight test sites in stratified unconsolidated glacial deposits of the north eastern USA.

3.3.3 Seismic reflection

Birkelo et al (1987) report the use of the common depth point method to measure the upper surface of a cone of depression developing during a pump test at a borehole some 25m from the survey spreads. Prior to pumping, the water table was about 3m deep and during the test (of 8 days duration) it was drawn down approximately 4m. The application of a low cut (600Hz) filter eliminated noise from the pump and hence the survey was continued while the pump was operating.

Ideal conditions for shallow reflection work include damp, fine-grained overburden and where such conditions exist the water table will generally be shallow and probably best detected by augering . In the simplest reflection technique (optimum offset) the water table is commonly used as the datum of the seismic traces.

3.3.4 TEM

Fitterman (1987) reports the application of TEM soundings (square loop with side 160m) to determine the water table depth and the thickness of saturated alluvial fill in a basin in New Mexico, USA. The sounding curves typically showed three layers: unsaturated alluvium underlain by saturated alluvium and finally a more resistive bottom layer (which in the present case is a tightly cemented conglomerate). The interpreted resistivities were, in descending order, 33ohm.m, 7ohm.m and 45ohm.m and the depth to the water table 129m, which matched closely the proved depth at 138m. The aquifer thickness was shown to be about 100m. The interpreted resistivity of the aquifer (7ohm.m) is very low; the resistivity of the contained groundwater was found to be 26ohm.m and the predicted resistivity of the aquifer (assuming 30% porosity) would be in the order of 250ohm.m. Fitterman concludes that the aquifer must contain a high proportion of clay minerals.

3.3.5 ABEM

The application of two frequency helicopter borne EM surveys to rapidly map the depth to the water table (between 25m and 100m subsurface) over large areas of western USA is described by Paterson and Bosschart (1987). This is one of few airborne geophysical surveys commissioned specifically for groundwater exploration; the authors believe that with increasing sophistication of airborne equipment and processing routines, such surveys will become more common.

3.3.6 GPR

Pullan et al (1994) observed the water table in coarse, resistive beds as a very strong GPR reflector at depths of about 7m while investigating an aquiferous moraine in Ontario, Canada. The technique would not be so useful in fine-grained environments where much of the electromagnetic energy would be absorbed.

3.3.7 EKS

In this most recent hydrogeophysical technique the water table is marked, in theory, as the point (in time after triggering) where the two voltage traces first move into and remain in phase while making significant excursions above the noise level. Data from shallow (2m to 4m) water table areas generally supports this supposition. However, where the water table is known to be deep such characteristic trace behaviour has frequently been observed at times (and hence depths) too soon to reflect the true water table; it is assumed that this condition indicates either a perched aquifer or the generation of EK signal in partially saturated sediments (see Peart et al, 1995, Beamish and Peart, 1996 and Peart et al, 1996).

3.3.8 Integrated techniques

Fitterman et al (1991) report that neither TCM nor TEM soundings were able to determine the water table in desert alluvium because the transition from partial- to full saturation is usually gradual and hence not well defined electrically.

VES and IP sounding

Sternberg et al (1990) report the use of combined resistivity/induced polarisation dipole-dipole sounding to determine very deep (c 150m) water table depths in alluvium of Arizona. The saturated alluvium is characterised by a moderate reduction in resistivity and a very large increase in chargeability. At one site the resistivity sounding yielded depths and specific resistivity values that were too high by a factor of 1.5; this was ascribed to anisotropy. The authors point out that Sumner (1976) has found consistently smaller intrinsic anisotropies of chargeability (IP) than resistivity, in the case of one sample in the ratio 1:44. Clearly this is a further advantage of performing combined resistivity/IP soundings.

3.4 DETERMINATION OF GROUNDWATER QUALITY

The ratio of fresh water to salt water conductivities may exceed three orders of magnitude and hence the electrical techniques offer a ready means of determining water quality. DC and EM profiling are used to identify the lateral extent of saline water while VES and EM sounding will yield information on the depth to the fresh/salt water interface and the thickness of fresh water lenses etc. Where an aquifer's porosity and clay content are known then fairly accurate estimates of salinity may be made on the basis of observed resistivity.

3.4.1 VES

Zohdy et al (1974) describe the successful mapping of fresh/salt water interfaces by the USGS using Schlumberger array VES. Al-Ruwaih and Ali (1986) used VES to map the potable/saline interface at about 30m depth in Pleistocene sediments of Kuwait. They found that the introduction of an intermediate layer of resistivity 10ohm.m resulted in an improved match between observed and modelled curves and it was possible to map variations in the thickness of this brackish layer across the survey area. The authors state that knowledge of such a layer is important for higher pumping rates can be sustained in zones of thick brackish transition layers since this suppresses upconing of truly saline water. Al-Ruwaih and Ali (op cit) also produced contour maps of both the aquifer resistivity and its transverse resistance. The first of these showed a strong inverse correlation with an isosalinity map based on borehole samples. Hence it could be assumed that the texture and composition of the aquifer was largely constant. The transverse resistance map indicated the most promising sites for drilling new boreholes (high values); low values indicate a thin aquifer, high clay content or saline water, or a combination of these factors. This is confirmed by the distribution of existing successful boreholes.

A VES survey to investigate groundwater quality and extent within an unconsolidated coastal plain aquifer in the Yemen is described by van Overmeeren (1989 b). Schlumberger array soundings were made at a density of about 1/6km^2 to a maximum current electrode separation of 1.8km. A simple qualitative examination of the resulting curves revealed four distinct hydrogeological environments: proceeding inland from the coast these are a) a 5km broad saline invasion zone with all observed apparent resistivities less than 1ohm.m, b) a transition zone (15km wide) of probably brackish groundwater (apparent resistivities about 3ohm.m), c) the potable aquifer, extending inland for some 35km (apparent resistivities of about 10ohm.m), and finally d) a zone of shallow basement. With the exception of the last zone, soundings in all remaining areas ended with a descending branch indicating a resistivity of less than 3ohm.m (probably reflecting a Tertiary formation containing saline water).

Unfortunately there was little additional information to incorporate into a quantitative interpretation of the soundings and, in an attempt to resolve the ambiguity resulting from the equivalence problem, the aquifer was modelled first with a fixed depth (ie assuming formation resistivity variations only) and secondly with fixed resistivity (assuming thickness variations of the aquifer only). Subsequent drilling showed the fixed thickness model to be most accurate. The formation resistivity variations required in this model were seen to be systematic and were subsequently related to the depositional processes involved ie. the higher resistivities in the west represented well sorted aeolian deposits (higher permeability) while a central belt of slightly reduced formation resistivity indicated alluvial (poorly sorted) deposits. This survey resulted in the discovery of a vast, formerly unknown, potable water resource.

Gondwe (1991) reports the application of VES to investigate possible saline invasion of a coastal alluvium and marine limestone aquifer in southern Tanzania. The optimum values of specific resistivity were found to be in the range 30ohm.m to 80ohm.m while values less than 10ohm.m were generally assumed to reflect either saline water or clay deposits. Lengthy geo-electrical profiles extending inland for some 10km indicated an assumed freshwater/saline interface at a fairly constant depth of 50m, contrary to the Ghyberg-Herzberg principle. The "interface" was subsequently found by drilling to correspond to a clay horizon. Elsewhere the low resistivity values were found to correspond to saline water. This ambiguity could have been resolved by the inclusion of a limited number of induced polarisation soundings.

Oatfield and Czarnecki (1991) describe combined hydrogeological, hydrochemical and geophysical surveys used to investigate the deep alluvial fill of the Amargosa Desert of Nevada, USA. They show how contours of the average transverse resistance (derived from VES curves) of layers interpreted in the uppermost 75m closely resemble the contour patterns of sodium concentrations as measured in borehole samples, there being an inverse relationship.

El-Waheidi et al (1992) studied a shallow alluvial aquifer in Jordan using deep Schlumberger array VES (AB to 2 000m). The equivalence problem was largely overcome by correlating proved lithological thicknesses at isolated boreholes with adjacent VES interpretations and extending the assumed fixed resistivity values to distant soundings. The authors report the successful delineation (laterally) of a saltwater/freshwater interface and moderate success with the identification of the depth of this interface.

3.4.2 Combined VES/IP

The differentiation between saline groundwater and clays/shales (both sources of low resistivity) has been achieved in a qualitative manner through the combined application of VES and IP soundings (Roy and Elliot, 1980). Saline groundwater is characterised by a simultaneous decrease in both resistivity and chargeability whereas clays demonstrate low resistivity but high (membrane) chargeability. The respective parameters may be readily observed qualitatively either on a joint sounding curve or profile. It should be noted, however, that this simple indication is not foolproof, for negative chargeabilities are occasionally observed in certain layered earth situations, even in the absence of low resistivity values.

Seara and Granda (1987) describe the application of joint VES/IP soundings to investigations of complex multi-layer coastal plain Miocene aquifers in the north east of Spain. Clay-rich horizons and those bearing saline water (resulting from over-exploitation) are readily differentiated following quantitative interpretation of the sounding curves and the construction of geo-electrical sections. The landward extent of saline intrusion is clearly defined by low chargeability and resistivity values and a close inverse correlation between weighted mean chargeability and sodium chloride concentration is noted.

3.4.3 TCM

Barker (1990) illustrates the cost-effectiveness of TCM using the Geonics EM34 system to delineate the lateral extent of saline intrusion in the shingle of Dungeness Spit , Kent, UK. This application had proved difficult for VES owing both to the extremely high resistivity (10 000 to 80 000ohm.m) of the coarse shingle at the surface (leading to contact resistance problems) and the steep descent of the resulting sounding curves which tends to suppress the signature of the important intermediate layer (the potable aquifer) (see Oteri, 1983) . Hence a survey was made using the EM34 system, using loop separations of 20m and 40m. Only the vertical coil configuration was considered as this is less sensitive to misalignment errors and the assumption of low induction numbers remains more valid in areas of high conductivities. A contour map of ground conductivities was prepared for each separation and these clearly show (qualitatively) the extent of saline intrusion.

3.4.4 TEM

Fitterman (1987) describes the use of TEM to map both the horizontal and vertical limits of landward transgression of sea water into an unconsolidated coastal aquifer in Massachusetts, USA; he also suggests the regular application of the technique to monitor temporal variations. A traverse of soundings (spaced at 600m interval) was run perpendicular to the coast; the inversion of the results was constrained by depths to the saline layer and bedrock proved in a nearby borehole. The interpreted geoelectrical section revealed a "wedge" of saline water, deepening and thinning landwards, underlying the potable aquifer. There was a distinct landward increase in the resistivity of the saline layer, probably reflecting the progressive dilution in this direction.

Fitterman and Stewart (1986) describe numerical modelling of situations similar to that just described; they investigate the limits of detectability of the saline layer and means of overcoming or minimising the ambiguities posed by equivalence. Their results underscore the concept that TEM is ideally suited to detecting conductive targets and offers very good vertical resolution.

TEM was used by Mills et al (1987) to map the landward advance of saline water in a multi-layer coastal alluvial aquifer with intervening clay layers in Monterey, USA. The upper aquiferous layer, approximately 50m deep, is the most heavily utilised and the saline intrusion in this has proceeded some 8km inland while in an underlying aquifer (at a depth of 120m) there is little invasion at present. In order to penetrate the upper, saline aquifer, a two stage programme of TEM sounding was employed: the initial measurements were made using loops of side 100m while information on the deeper aquifer was derived from loops of side 200m. By careful calibration of the sounding data at numerous test wells where the water quality was well known, it was possible to map the position of the 500mg/L isochlor (equated with a value of 8ohm.m) in both the upper and lower aquifers. Additional details of the interpretation of this survey together with an excellent account of the field procedures for undertaking TEM soundings are to be found in Hoekstra and Blohm (1990).

3.4.5 ABEM

Three frequency ABEM data (fixed wing) acquired over a strip of Kenya's coastal plain were converted to a simple one layer model showing apparent ground conductivity (Paterson and Bosschart, 1987). The resulting map clearly delineates a 2km broad zone of saline intrusion (high conductivity), a belt of very low conductivity (reflecting a landward belt of coarse sandy soil, a zone of high recharge) and finally, a zone of intermediate conductivity that reflects near surface, impermeable shales.

Oteri (1991) describes the application of an INPUT survey with galvanic resistivity follow-up to outline an area of shallow potable water in Kenya's coastal Cainozoic aquifer (comprising sands, corals and clays). The contoured fifth-channel INPUT data delineated an area of relatively low conductivity along Kenya's coastal strip, measuring approximately 4km by 1km. Surface resistivity surveys confirmed shallow high resistivity values (up to 10 times background) over this area which was interpreted as representing a fresh-water lens. Analysis of water from shallow handpumped boreholes here confirm this interpretation. Oteri (op cit) notes that freshwater is also obtained from zones displaying resistivity values as low as 10ohm.m; such low values are attributed to a high clay content.

3.4.6 Combined TCM/VES

Van Overmeeren (1989 a) describes a combined approach comprising rapid and cheap EM34 profiling with occasional VES at carefully selected sites to delineate in both the lateral and vertical sense, freshwater bearing sandy creekbeds in a mainly saline coastal area of the Netherlands. The survey procedure involved the measurement of conductivity with all six possible combinations of coil separation and orientation over an area of some

800km^2 at a station density of 1/km^2. An average of 40 stations were observed each day by a two-man crew. The data were then converted to values of apparent resistivity and six contour maps produced. The shallowest penetration map (vertical coils at 10m separation) gave the best indication of the lateral extent of the narrow, sinuous, freshwater bearing channelways while the changes between the map sets indicated changes with depth and hence yielded an indication of depth extent of the potable aquifers. VES were made at selected sites, mainly within the freshwater areas but including a few test sites where the TCM data indicated saline conditions. The VES data generally confirmed the TCM indications and yielded quantitative data on aquifer resistivity (between 30ohm.m and 70ohm.m) and thickness (up to 25m).

Van Overmeeren (op cit) attempted quantitative interpretation of the TCM data but recognised two of the limitations of TCM in this application ie. the technique responds best to conductive targets underlying resistive features (the converse of the present situation) and the saline (high conductivity) environment breached the instrumental constraint of operation at low induction numbers. Thus he attempted to derive empirical relationships between TCM indications at various penetration depths and the observed aquifer characteristics. Van Overmeeren concluded that TCM-mapped values of 25ohm.m or higher indicated freshwater bearing aquifers whose thickness exceeded 10m, whereas values lower than 10ohm.m implied very saline water less than 4m below surface.

3.5 MISCELLANEOUS APPLICATIONS OF SURFACE GEOPHYSICAL METHODS

3.5.1 Aquifer properties

The classical methods of determining aquifer properties (including pump testing, ponding experiments and laboratory measurements on samples (which may not be truly representative of the entire aquifer) are both time consuming and expensive. Fortunately several aquifer properties (porosity, permeability and grain size) may be estimated satisfactorily on the basis of geophysical parameters that are derived almost as a bi-product of routine surveying. Such estimates have the added advantage that they reflect bulk values. Intuitively the electrical methods are especially suitable for determining aquifer properties because both water and electric current tend to follow similar paths through a rock matrix (ie the path of least resistance). Both seismic velocity and density parameters can also yield useful information on aquifer characteristics.

3.5.1.1 Porosity

Ayers (1989 a) describes the derivation of aquifer porosity from seismic velocity. Using the empirical Nafe and Drake (1963) relationship between compressional wave velocity and bulk density, the velocities are first converted to densities and these, in turn, are converted to porosities through the equation (Freeze and Cherry, 1979):

$$\Phi = 1 - (\rho b/\rho g)$$

where Φ is porosity, ρb is bulk density and ρg is grain density (quartz). Ayers (op cit) points out that the geophysically derived porosities in this particular study are generally higher than values obtained during other hydrogeologic studies. This probably results from the fact that the lowest velocities represent increasing content of silts and clays which, while reducing the predicted density, will reduce the effective porosity. Provided the error is systematic it can, of course, be corrected for each different study area.

Ayers (1989 b) also derives porosity from formation resistivity values resulting from electrical sounding interpretation. By applying the Archie equation modified for unconsolidated sands:

$$F = 1/\Phi^{1.3} = p^b/p^w$$

where F is the formation factor, Φ is porosity, p^b is bulk formation resistivity (from VES data) and p^w is the pore water resistivity measured from local well samples. The porosity estimates for the sand/gravel aquifer of the Lower Platte system range between 0.27 and 0.43, which values agree well with those derived from seismic velocity and described above (Ayers, 1989 a).

3.5.1.2 Permeability or Hydraulic conductivity

Following Archie's (1942, 1950) early work linking formation factor, porosity, permeability and subsequently hydraulic conductivity, numerous investigators have reported either direct or (more rarely) inverse relationships between hydraulic conductivity and the resistivity of granular aquifers; it appears that the precise nature of this relationship is governed primarily by the average grain size and the degree of sorting of the aquifer.

Heigold et al (1979) made VES at 5 pump-tested boreholes along the axis of a glacial outwash aquifer in Illinois, USA, where hydraulic conductivity values were available. The electrical conductivity of water samples collected from 2 widely space locations throughout the survey area remained fairly constant and hence they assumed that the primary control on aquifer resistivity was hydraulic conductivity. Using data from only three of the boreholes, Heigold et al (op cit) observed an inverse relationship between hydraulic conductivity and aquifer resistivity. In the attempt to explain this controversial finding, the authors undertook sieve analyses on aquifer samples and concluded that their results reflected differences in the sorting of the outwash sediments forming the aquifer (ie more poorly sorted sediments are responsible not only for reduced porosity (and hence lower hydraulic conductivity) but also for an increase in the volume of low conductivity solids which increase the resistivity of the aquifer). It should be added too, that their results based on observations at only three boreholes, all displaying comparable hydraulic conductivities, are at best of dubious value.

Van Zijl et al (1981) report a correlation between the resistivity of water saturated primarily sandy, alluvium (corrected for water quality) and permeability and also between porosity and permeability.

Allessandrello and Lemoine (1983) summarise 20 years' observations in the alluvial plains of the Saone, Rhone and Loire Valleys of France and demonstrate that a precise (but site specific) direct correlation exists between permeability and formation factor (ie the resistivity of the formation divided by the groundwater resistivity) for each environment, provided that variations in groundwater conductivity are accounted for. Since aquifer thickness may also be derived from a VES curve, then maps of transmissivity (the product permeability * thickness) may be derived readily from a series of VES. The authors appear to have largely ignored the influence of silts and clays; possibly the aquifers they examined are particularly well sorted and hence clean.

Ponzini et al (1984) examined the relationship between transverse resistance (corrected for variations in pore water quality) and hydraulic transmissivity of an unconsolidated alluvial (sand and gravel) aquifer sandwiched between two clay horizons, deposited in the Tronto River valley of eastern Italy. The authors point out that transverse resistance is an especially useful parameter since it can be derived uniquely from a VES curve, ie it is not subject to the problem of equivalence. They also state that the derivation of hydraulic conductivity from transmissivity data obtained during pump tests is also liable to errors and hence it is especially expedient to seek a correlation between transverse resistance and hydraulic transmissivity. But first the computed transverse resistance values must be normalised by dividing by the porewater resistivity, to account for variations in this parameter. A further correction is required in those cases where the aquifer is phreatic and the unsaturated/saturated boundary cannot be resolved on the VES curve: it is essential to calculate the component of transverse resistance due to the unsaturated layer (which may well be dominant) by independent means. The authors demonstrate a very strong direct but non-linear correlation between the corrected transverse resistance and hydraulic transmissivity that is also shown to be valid for two additional and similar Italian alluvial aquifers. They stress, however, that this particular relationship cannot be extended to aquifers deposited under different conditions for which a specific relationship should be established. Such correlations are only likely to be strong in the case of simple (non-layered) alluvial aquifers of fairly homogeneous composition and hence uniform porosity.

Niwas and Singhal (1985) extended the original direct relationship they observed between transverse resistance and aquifer transmissivity to account for variations in water quality. They demonstrated the applicability of their revised equation in studies of a glacial outwash aquifer of Rhode Island, USA and in three alluvial aquifers in India, achieving encouraging results in all four areas. Razack and Sinan (1988) tested several statistical regressions (linear, exponential and geometrical) of hydraulic transmissivity on transverse resistance for a strongly heterogeneous alluvial aquifer on the Haouz Plain of Morocco. They found a reliable linear geometrical regression of the two parameters when the permeable beds were treated in isolation.

Mazac et al (1990) found both direct and inverse relationships between hydraulic conductivities and rock resistivities (as defined by VES) of saturated porous aquifers. The direct relationship exists between the main rock types (characterised by grain size); thus gravel (at the coarse end of the scale) displays both high resistivity and high hydraulic conductivity whereas a sandy-clay (towards the fine end of the scale) displays reduced resistivity and hydraulic conductivity. However, an inverse relationship is observed within any particular rock type (characterised by a certain grain size) that is accounted for by porosity variations. Due to anisotropy, the precise nature of these relationships is modified by the direction of groundwater flow (vertical or horizontal), the type of bedding and the type of resistivity measurement (longitudinal or transverse) observed. Mazac et al (op cit) undertook experiments in saturated sandy gravels of a river in southern Bohemia (Czech Republic) and demonstrated a direct relationship between hydraulic conductivity (K) and aquifer resistivity (p) of the form:

$$K\ (10^{-5} m/s) = 97.5^{-1} * p^{1.195}$$

The authors stress that while their general conclusions should be universally applicable, the particular equations must be calculated afresh for different hydrogeological environments.

EK surveying yields an indication of the variation of permeability with depth (Beamish and Peart, 1996). Laboratory experiments with fluid saturated porous rock have shown that the rise-time of electrokinetic voltages reflects directly the permeability of the rock. This result has been extended to field EK surveys so that an indication of the variation of permeability with depth is derived from the observed voltage/time data. The approach adopted requires estimates of the bulk moduli of the fluid and solid constituents, shear modulus of the solid frame and porosity. Although porosity may be iteratively adjusted during estimation of permeability, appropriate elastic moduli must be assigned in each new survey area. It should be noted therefore that the current estimates of EK permeability should be regarded as relative rather than absolute determinations.

3.5.1.3 Aquifer grain size

Haeni (1995) describes the correlation between interpreted electrical resistivity, derived from VES and VLF-R data, and grain size at eight sites in stratified unconsolidated glacial drift of the north eastern USA. He notes that where the specific electrical conductance of the groundwater remains relatively constant, the bulk resistivity of the aquifer is generally representative of the aquifer's grain size characteristics, coarse grained material being more resistive than fine grained material. Haeni (op cit) states that the EM techniques are better able than DC-VES to resolve small scale changes in electrical properties.

3.5.2 Groundwater volume and aquifer storage change

West and Sumner (1972) describe how the volume of groundwater contained in a basin may be determined by gravity mapping. The mass deficiency observed over a basin reflects the density contrast between bedrock and the contained saturated and unsaturated fill and the total pore volume. If an accurate residual gravity map is available, then by the application of Gauss's law and by including values for the average density and porosity of fill and the average storage coefficient (all of which can either be estimated or obtained from boreholes), then the groundwater volume may be derived. The authors state that this is a rapid and cheap method for making at least a first order estimate of the groundwater storage capacity of a basin.

Pool and Eychaner (1995) report the use of repeated gravity observations (temporal gravity surveys) to estimate aquifer storage change and the specific yield of unconfined aquifers where significant variations in water levels occur. The method was first suggested in 1974 but it is only in recent times that sufficiently sensitive gravity meters (such as the LaCoste and Romberg Model D) have become available with which to observe the small gravity changes involved. The relative difference in observed gravity is measured between reference stations on stable bedrock remote (at least 10km) from the aquifer (where mass changes are expected to be minimal) and stations over the aquifer. The survey is repeated after a period of time to establish changes in the differences in observed gravity. In the present work gravity changes at borehole stations ranged from 77microgals to 158 microgals (standard deviation 1.6microgals to 5.9 microgals) whereas changes at the bedrock reference stations varied in the range 6 microgals +/- 3 microgals. The smallest change in aquifer observations were near the sites of perennial streams (ie where water table fluctuations were smallest). The authors conclude that the values of aquifer storage and specific yield derived by this method were similar to those derived from aquifer-test analyses; a major advantage of this technique results from the much larger aquifer volume sampled.

3.5.3 Contamination studies

Alluvial aquifers are especially liable to contamination because not only are they generally shallow but they also occupy the typically favoured sites for urbanisation and industrialisation (ie broad, flatlands). Their typical complexity of layering and lithological variety may render the occasional monitoring borehole almost useless, besides which, it is poor practice to penetrate protective layers by drilling where contaminants are involved; for this reason any rapid surface technique that can supply detailed information on intra-alluvial features and the extent of contamination (and hence guide the siting of test holes) is particularly valuable. Since minute quantities of both inorganic and organic contaminants will affect groundwater conductivity strongly it is not surprising that the electrical techniques are employed almost exclusively in contamination studies.

3.5.3.1 DC Resistivity

Urish (1983) reports a galvanic resistivity survey (both sounding and profiling modes) to map contamination in a buried glacial valley infilled with alternating fine- and coarse grained sands. He emphasises the importance of making initial soundings in order to determine the most suitable (sensitive) electrode separation for subsequent profiling; the author also makes useful comments on the practicalities of running successful resistivity surveys.

Mazac et al (1989) describe a combined surface-borehole galvanic resistivity method (the screening body technique - a form of mise-a-la-masse) and conventional resistivity profiling to delineate the extent of pollution by discarded oil-products in a sand and gravel aquifer in Czechoslovakia. In the screening body technique, one current electrode is placed in a borehole immediately below the polluted layer while the other is placed at "infinity" (typically about 1km away). The resulting potential field is mapped on the surface by a roving pair of potential electrodes and meter. In the example given, it is clear that this technique has far greater detection potential than the purely surface traverses.

Osiensky and Donaldson (1994) describe a further modified mise-a-la-masse resistivity system to monitor the evolution of a conductive tracer plume migrating through fluvial sands and gravels. One current electrode was planted in a ditch excavated by back hoe that was used to introduce the tracer to the aquifer; the remote current electrode was placed some 200m distant. They then observed at regular time intervals the equipotential patterns developed across a rectangular grid of about 200 non-polarising electrodes covering the study site. The evolution of the plume with time (and hence the preferred pathways) was clearly depicted by the equipotential pattern.

Auken et al (1994) describe the application in Quaternary deposits of Denmark of a vehicle-towed shallow resistivity measuring technique known as pulled array continuous electric profiling (PA-CEP). The system is micro-processor controlled and data is obtained for up to three electrode spacings/arrays at horizontal intervals of about 1m. Such data allows high resolution resistivity mapping of the upper 15m or so. The authors rapidly acquired 100km of resistivity profiles and were able to outline vulnerable zones where the clayey moraine protective cover overlying aquiferous glaciofluvial deposits was thin or absent. Sorensen (1994) describes the pulled array continuous electrical profiling system in some detail; he also demonstrates the high degree of repeatability of this system when compared with a 300m traverse acquired using conventional Wenner array traversing. Christensen and Sorensen (1994) report that work is in hand to produce a PA-CEP system that will measure 10 electrode separations and this will effectively provide a complete electrical sounding at 1m intervals.

Contamination of a coastal freshwater-lens type aquifer, of well sorted sands and glacial sediments, resulting from seepage from a landfill site near Cape Cod, USA, was detected by Frohlich et al (1994) following a survey comprising a traverse of Schlumberger array VES and subsequent Wenner array profiling. The resulting VES curves were all steeply descending, reflecting dry, high resistivity surface sands (up to 10 000ohm.m) underlain at depth by salt-water saturated sands (8ohm.m). All the VES curves also displayed a central inflexion which represented the freshwater zone. Careful curve matching showed that at three adjacent sounding sites this inflexion implied the presence of two layers of intermediate resistivity, the lower layer (of lower resistivity) indicating the polluted freshwater. Wenner array profiling (with electrode separation chosen to "focus" on the freshwater lens) confirmed the presence of a belt of low resistivity some 140m broad that was subsequently seen in monitoring wells to delineate the contaminant plume.

3.5.3.2 TCM

McNeill (1989) describes the mapping of an industrial contaminant plume in a shallow alluvial series underlain by a clay aquitard using EM conductivity equipment (Geonics EM31 and EM34). The survey involved traversing with both sets of apparatus along a 1km transect with shallow control boreholes at 70m centres. The lateral extent of the main plume was clearly defined by conductivity peaks on all traverses, while a second plume was indicated on the wider coil separation traverses. The amplitude variations at the different coil separations enabled estimates to be made of the vertical extent of both plumes. The indications were confirmed by subsequent induction logging and water sampling in the control boreholes. McNeill (op cit) points out that the conductivity traverses can be made so rapidly that the technique is very valuable for monitoring, by repeated measurements, the advance of the plume with time. It would also be possible to make selected 1-D interpretations of the different coil responses to obtain a more quantitative estimate of the vertical extent of contamination.

Goldstein et al (1990) describe combined EM31 and EM34 surveys to map the extent of a contaminant plume at a depth of between 5m and 20m under saline soils. Instead of solving numerically for the layered-earth inverse solution that fits the observed data at each station, the authors applied a more simple technique to enhance the response to deeper conductivities. Their procedure is to calculate a new apparent conductivity at each station according to the formula:

S (new) = 2*s(40m) - s(20m)

These values are plotted to yield a ground conductivity map that emphasises conditions between about 8m to 16m deep. Combining this map with the EM31 data it was possible to delineate the position of the (deep) contaminant plume and various additional shallow anomalously conductive features.

3.5.3.3 GPR

The capacity of GPR to distinguish small scale changes in stratigraphy and physical properties in an apparently homogenous sandy aquifer at the Borden Test Site in Canada are highlighted by Bauman et al (1992). Using a frequency of 100MHz they were able to define very subtle stratigraphic features (including high- and low angle bedding), probably due to the variable water content of individual layers, and these are believed to have a controlling influence on the migration of spilled solvents. The authors point out that the principal limitation of the method is the maximum depth of penetration of only some 6m due to conductive attenuation. Deeper penetration could be achieved with lower frequencies but this would result in a loss of vertical resolution.

3.5.3.4 Integrated techniques

Greenhouse et al (1989) describe successful contaminant surveys in alluvial aquifers using various combinations of TCM (Geonics EM31 and EM34), GPR (Pulse EKKO IV), VLF(R) and galvanic resistivity (dipole-dipole array). They stress that in mapping contaminants it is sometimes difficult to distinguish between the signal (ie the physical response to the presence of the contaminant) and geological noise (resulting from the presence of conductive clay lenses, variable moisture content and topographic effects etc), especially where the contaminants are in low concentration. However, the expected plume-like appearance of an anomaly, where anomalous values become progressively weaker extending away from the source, has been found to be a powerful discriminator. Such anomalies are enhanced when the data is normalised by the background response and/or expressed in logarithmic units. The authors also describe a simple topographic correction for apparent conductivity values whereby all the values recorded are plotted against the respective station elevation to yield a function that can be applied to normalise for elevation differences. Of course, these "noise" problems do not arise in the case of contamination monitoring.

Greenhouse et al (op cit) show the results of surveys to map a contaminant plume that emanates from a landfill site in Ontario and flows through glaciolacustrine sands overlying irregular Precambrian bedrock. The sands saturated with natural groundwater recorded a fairly uniform conductivity of 2.5mS/m and this value was used to normalise the results of both VLF(R) and EM31 surveys which both revealed the plume as positive contour levels whereas shallow bedrock appeared as negative values. A subsequent GPR traverse, normal to the plume, revealed a pattern of signal attenuation that correlated almost precisely with groundwater conductivities in excess of 100mS/m (as measured following an intensive drilling programme). Thus the GPR survey (in this favourable environment) provided valuable additional information on the vertical structure of the plume.

Brune and Doolittle (1990) describe TCM and GPR surveys using the Geonics EM34 and a vehicle-towed (3km/h to 5km/hr) continuous recording SIR-8 GPR system, respectively, to locate seepages of animal wastes from lagoons at isolated sites where the sealing had failed. The leakage sites were clearly identified by plumes of high conductivity (up to 16 times background) extending away from the lagoon walls; these same features were recorded on the radargrams as zones of severely attenuated signal and/or dramatic change of radar signature. The authors note, however, that the penetration of the GPR signal is limited to some 2m only in zones of clay. Subsequent sampling of shallow (to 12m depth) groundwaters within and outside of the plumes proved relatively high content of chloride and total ammonia. The authors suggest that it may eventually be possible to quantify the concentration of pollutants on the basis of the TCM responses.

Additional PA-CEP surveys, integrated with VES and TEM soundings, to determine the vulnerability to pollution of some of Denmark's glacial-derived aquifers, are described by Christensen and Sorensen (1994). Quoted production figures for the PA-CEP, with two operators, are between 10km/day and 15km/day using three discrete electrode separations and with an observation interval of only 1m. This is at least 7 times the rate for conventional galvanic coverage. Large areas are rapidly covered at high data density. The data for each electrode separation are plotted as iso-ohm maps. These highlight zones of high resistivity (likely to indicate sand and hence vulnerability) while the changes between the maps achieved at different separations indicate the depth extent of the sands etc. The authors also point out that the iso-ohm maps can be used to best site subsequent VES which should be made only in zones displaying little lateral variability.

Matias et al (1994) report the rapid mapping using Geonics EM34 of a contaminant plume in 30m thick highly porous and permeable Holocene sediments in Portugal. The EM data outlined a zone of high conductivity (3 times background) immediately beyond a landfill site and extending over 400m from this. They subsequently made VES on two traverses crossing this feature; these confirmed the lateral position of the plume and indicated its depth extent.

3.6 Predictive modelling

The ready availability of both powerful personal computers and software packages for the interpretation and modelling of geophysical data has facilitated the procedure of predictive modelling. Thus it is now possible to model the response to be expected by most geophysical techniques for a range of targets and environments. Such modelling should be an essential preliminary step in any exploration programme; not only will it indicate the feasibility of the survey (thereby helping to avoid fruitless field trips etc), it can also show the optimum observation interval, array etc to be used and thus the survey can be made more efficient.

It is suggested that the hydrogeologist who is faced with a particular problem concerning the exploitation of UNSAs should first consult the relevant summary sheets (following) and decide which technique or combination of techniques are likely to be most suitable in the conditions anticipated. The final choice of techniques should be made following comparative forward modelling to ensure that measurable responses will be yielded by targets of likely dimensions situated in realistic environmental conditions. It may well be decided at this stage that a geophysical survey is not likely to be successful.

Haeni (1995) has produced an excellent account of the application of four popular geophysical techniques for the investigation of glacial sand and gravel aquifers in north east USA. His work includes the forward modelling of VLF-R and TCM (Geonics EM34) responses to be expected over various hypothetical but realistic geoelectrical earth models (eg coarse grained aquifer material overlying fine grained material). On the basis of these studies Haeni (op cit) demonstrated that both of these EM techniques are able to detect subtle horizontal and vertical electrical changes in the subsurface but that each technique has a preferred environment. For instance, VLF-R is most effective in shallow resistive conditions while TCM performs better in conductive environments.

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6. SUMMARIES OF SUGGESTED PROCEDURES AND KEY REFERENCES

6.1 DETERMINATION OF BEDROCK TOPOGRAPHY AND THICKNESS OF UNCONSOLIDATED SEDIMENTS

In these applications geophysical surveys have been undertaken at all scales, from the regional investigation of major sedimentary basins to the very local tracing of buried valleys incised into bedrock and subsequently infilled with alluvial deposits.

Basin investigations

Gravity is recommended where there exists a large density contrast (say greater than $0.25g/c^3$) between unconsolidated basin-fill and bedrock and where a sediment thickness of at least 50m is anticipated. Ideally the area to be surveyed should have low relief and be remote from mountains. Suitable coverage for large scale investigations may already exist (eg as part of a nation-wide survey).

Aim for an observation density of about 1 station/km^2, traversing along convenient roads/tracks. Alternatively, work on straight profiles (normal to the basin's major axis where known), making observations at an interval of about 250m. Station elevations can be determined either from bench marks or contour intersections, by using barometric altimeters or, in flat-lying areas, can be ignored. With modern, low-drift gravimeters, it may not be necessary to return to a base station every 4 hours or so (since tidal corrections can be determined from published tables). Numerous corrections (for latitude and elevation variations etc) are applied using a PC. The traditional, onerous corrections may not all be necessary, depending on the scale of anomaly anticipated and the degree of precision required **(Angelillo et al (1991))**.

Following data reduction either a Bouguer anomaly map or a series of profiles are produced. The regional effect should be removed (by trend surface analysis) so that the remaining residual data reflects only local variations in the depth to bedrock. A qualitative assessment at this stage will indicate the outline of the basin and zones of deep/shallow bedrock. 3-D modelling typically assumes a single density contrast and iteratively adjusts the bedrock topography (as a series of square prisms) until an acceptable match between observed- and calculated anomaly is achieved. 2-D modelling (assuming infinite strike extent) is much quicker; it is undertaken on profiles measured in the field or abstracted from contour maps. Density values for modelling can be obtained by a **Nettleton (1939)** traverse (for unconsolidated deposits) and borehole samples (for bedrock). It is also possible to model without assuming densities by correlating residual gravity with bedrock depths as proved in boreholes. Software for data reduction and 3-D and 2-D modelling is readily and cheaply available.**(Birch (1982) and Ali and Whiteley (1981))**

Gravity is a rapid (100 observations/day), straightforward, single operator technique. Gravimeters are expensive (£30 000 to £40 000) but they can also be rented for about £50/day.

Seismic refraction using a sledge hammer and plate source is practicable for investigating bedrock down to about 30m; this limit can be extended to about 50m where explosives/falling weight sources are available. Investigations beyond this depth require excessively long spread lengths (as a rule of thumb, 10 times the depth of penetration) and large amounts of explosive. Ideal conditions comprise a strong velocity contrast between bedrock and sediments (at least 2:1), a shallow water table (hence a reduced section of seismically "lossy" unsaturated sediments) and homogeneity of the sediments (ie an absence of high velocity clay bands, thick calcretes, shallow perched water tables etc). Given such conditions, interpreted depth to bedrock should be accurate to better than 10%.

About 4 "spreads" can be completed (shooting forward and reverse) by a 3-man team in one day; the actual coverage achieved depends on the geophone spacing but typically would be between 500m and 2 000m. The simple time-intercept interpretation yields the depth to bedrock only at each shot point. More sophisticated reciprocal time methods give a depth to bedrock under each geophone. The Generalised Reciprocal Method of interpretation also indicates the depth at each geophone and in addition reveals lateral variation of bedrock velocity and enables hidden layer- and velocity inversion conditions to be recognised **(Palmer (1980), Haeni (1986), Ayers (1988) and Haeni (1995))**

The highest resolution of bedrock topography is provided by seismic reflection **(Pullan and Hunter (1990) and Roberts et al (1992))**. However, such surveys, requiring high frequency geophones, a high frequency source (hunting rifle etc) and special processing software, are not yet undertaken routinely. The technique is not suitable for very shallow work (say less than 5m sub-surface) nor does it yield good results in dry, unconsolidated (seismically lossy) material. In ideal conditions (shallow water table and large acoustic impedence at the base of the sediments) penetration up to 300m has been achieved.

VES surveys are frequently applied to determine depth to bedrock, exploiting the typically large resistivity contrast between saturated sediments and bedrock. Advantages include the cheapness (and hence widespread availability) of the equipment, simple field operation and the fact that the electrical techniques provide additional information on groundwater quality. **De Beer et al (1981)** quote cases where VES were applied because neither gravity nor seismic refraction techniques were suitable. Gravity failed because large density contrasts in the basement obscured anomalies reflecting the variations of alluvial thickness while velocity inversions resulting from occasional cemented layers precluded seismic refraction.

Optimum field conditions for VES include low relief, sparse vegetation cover and the absence of dry overburden (which causes contact resistance problems). The technique is best confined to the investigation of depths less than about 50m; the wide electrode separations required to investigate deeper than this result in a rapid loss of resolution as progressively larger volumes of ground are sampled. The DC resistivity techniques are slow compared to EM (non-contacting) methods; about 10 soundings can be undertaken (to a maximum electrode separation of about 500m) by a three man crew in one day. VES interpretations are subject to the pitfalls of equivalence and suppression and should be calibrated with borehole data where possible. In the absence of such data the mapped values of total longitudinal conductance (read directly from ascending VES curves) provide a reliable guide to bedrock depth.

EM sounding techniques are logistically superior to DC; TEM can provide better resolution at greater depths than VES and will certainly yield more accurate depths to bedrock where this is not highly resistive (eg consolidated shales) **(Fitterman (1987) and Auken et al (1994))**. Penetration of 150m to 200m is readily achieved using loops of side 100m and some 25 soundings can be completed in one day.

Occasionally complementary EM and DC soundings are made to help resolve the problems of equivalence **(Van Kuijk et al (1985))**. **Fitterman et al (1988)** have examined the equivalence and suppression displayed by the three sounding techniques (VES, TEM and TCM) over typical alluvium filled basins and demonstrate the situations where complementary application is indicated.

<u>Delineation of buried valleys</u>

Gravity observations should be made on profiles normal to the expected valley trend at an interval of about 25m. Optical levelling will be required (accurate to about 5cm) since the anticipated anomaly will be small (probably less than 1mgal). The estimated accuracy of a standard field survey is about 0.1mgal. Modelling may not be necessary where a distinct (usually negative) low amplitude anomaly can be traced from profile to profile. **(Carmichael and Henry (1977), Ibrahim and Hinze (1972) and Culek and Palmer (1987))**

The standard seismic refraction technique as described above can also be used to define buried valleys **(Haeni (1986)**. Such features have also been traced successfully (and occasionally more efficiently) by seismic "fan shooting"; this involves the progressive location of a slow velocity zone using an arcuate arrangement of geophones **(Kearey and Brookes (1984)**. Seismic reflection surveys are well able to map buried valleys **(Miller et al (1989))** but, as described above, the technique is not yet applied routinely.

Both VES and resistivity traversing can be used to locate buried valleys **(Zohdy et al (1974), Frohlich (1974) and Worthington and Griffiths (1975))**. The electrode spacing used for traversing should be determined following VES to ensure focusing in the depth of interest. Complementary IP traverses have occasionally yielded chargeability anomalies over valley fill that failed to respond to resistivity **(Ogilvy (1970))**.

TCM and HLEM traversing is much quicker than DC resistivity and both techniques have been employed to detect buried valleys **(Cornwell (1985), Jones (1986) and Beeson and Jones (1988))**. Variations in conductive overburden thickness in the top 30m or so have been also been mapped rapidly using VLF-R **(Poddar and Rathor (1983))**.

GPR has performed well in determining buried topography down to about 20m in resistive environments **(Davis and Annan (1989))**.

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6. SUMMARIES OF SUGGESTED PROCEDURES AND KEY REFERENCES

6.2 DETERMINATION OF STRATIGRAPHY AND COMPOSITION OF UNSAs

In hydrogeological modelling it is frequently required to determine the continuity of a particular horizon (an impermeable clay or aquiferous gravel, for instance). Similarly important is information on the compositional variations within an alluvial sequence.

Stratigraphy

The finest bed resolution (to 0.5m) coupled with ample penetration (to 300m) is provided by seismic reflection **(Pullan and Hunter (1990), Slaine et al (1990), Geissler (1989) and Meekes et al (1990))**. However such surveys, requiring high frequency geophones, a high frequency source (hunting rifle etc) and special processing software, are not yet undertaken routinely. The ability to resolve relatively slow velocity beds underlying high velocity layers is a further important advantage of seismic reflection. The technique is not suitable for very shallow work (say less than 5m sub-surface) nor does it yield good results in dry, unconsolidated (seismically lossy) material.

GPR does perform well in areas of dry, unconsolidated overburden and can provide high bed resolution and penetrate to about 20m **(Davis and Annan (1989))**. GPR responds to minor textural and moisture variations and hence the disposition of very thin layers may be defined. The technique fails to penetrate thin (1m) conductive horizons. GPR is rapidly becoming popular for shallow investigations. Progress is fast (up to 20km/day). The equipment is moderately priced (£30 000) or may be rented at about £1000/week.

VES can map discrete layers, preferably where these are relatively resistive, provided they display sufficient resistivity contrast with their enclosing lithologies and are not too thin in relation to their depth of burial (thickness at least one tenth of burial depth) **(Ayers (1989 b) and Van Zijl et al (1981))**. **Worthington (1977)** successfully mapped a deep and thick aquiferous unit of the Kalahari Deposits on the basis of its transverse resistance that could be uniquely determined from VES curves. **Tucci and Pool (1985)** delineated zones of fine-grained sediments (relatively low resistivity) and determined their thickness using VES. The technique is best confined to shallow investigations (depths less than 50m) due to the rapid loss of resolution with the large electrode separations required for deeper penetration. Galvanic contact with the ground is required; this results in slow progress and contact resistance problems may be encountered in dry overburden. About 10 soundings (with electrode separation of 500m) can be made by a crew of at least three men in one day.

TEM is logistically superior to VES and is especially useful for resolving and tracing relatively conductive units at depths down to 300m **(Fitterman et al (1991) and Taylor et al (1991)**. About 20 TEM soundings can be made in one day, again employing a crew of at least three men.

Joint TEM/VES was required to map both resistive and conductive horizons (and to overcome equivalence problems); TEM responded well to conductive layers and mapped the base of the alluvium (a conductive shale) while VES resolved the high resistivity horizons (sand and gravel) **Christensen and Sorensen (1994)**.

<u>Composition/content</u>

Shallow (less than 40m) compositional variation is best detected by TCM (EM31 and EM34); **Zalasiewicz et al (1985) and Potts (1990)** describe the ready differentiation of alluvium into gravel/sand/clay components on the basis of conductivity observations. **Hazell et al (1988)** used TCM traverses (sited using air photographs) with confirmatory VES to locate thick sands/gravels in riverine alluvium. TCM progress is rapid (10km/day with single operator EM31 (offering 5m penetration) and up to 5km/day with two-man EM34 (penetration to 50m).

Similarly, coarse sands and gravels enclosed in fine (conductive) sediments have been outlined using HLEM **(Verma and Bischoff (1989) and Balmer et al (1991))**. This is a non-contacting and hence very rapid technique with productivity similar to TCM). **Muller (1992)** employed multi-frequency VLF_R to map tortuous sands and gravel deposits enclosed in silts and clays; the use of three frequencies allowed some discrimination of depth extent. VLF is a passive, non-contacting, rapid technique giving penetration of some 30m.

IP traversing/sounding can be applied to locate "dirty" (clay rich) aquifers which display high chargeabilities due to membrane polarisation **(Vacquier et al (1957) and Bodmer et al (1968))**. The technique is relatively slow and the equipment expensive and hence not commonly available.

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6. SUMMARIES OF SUGGESTED PROCEDURES AND KEY REFERENCES

6.3 DETERMINATION OF WATER TABLE DEPTH

The probability of success of any technique in this application is broadly governed by the average aquifer grain size. In coarse sediments the water table will be readily defined as a sharp physical property boundary but in fine sediments a broad capillary fringe will develop, across which the physical properties will vary continuously, and accurate detection is unlikely.

VES has been applied traditionally and successfully, especially to shallow problems (less than about 40m) where the saturated layer has substantial thickness (say at least one fifth of the depth to the water table) and the underlying rock is of strongly contrasting resistivity. **(Haeni (1995) and Al-Ruwaih and Ali (1986))**. Galvanic contact with the ground is required and hence productivity is relatively slow (about 10 soundings per day with 500m electrode separation) and contact resistance problems will be encountered in dry overburden. The technique rapidly looses its resolution with depth due to the large volumes of ground sampled at wide electrode spacings.

Sternberg et al (1990) employed combined VES/IP soundings (dipole-dipole array) to detect a 150m deep water table; the saturated alluvium was characterised by a moderate reduction in resistivity but a very large increase in chargeability and the water table was therefore best defined by IP. However, such joint soundings are slow and IP equipment is not always available.

The water table is usually an ideal target for TEM since the saturated zone represents a conductive layer. **Fitterman (1987)** has accurately detected the water table at a depth of about 130m where it presented a resistivity contrast of about 5:1. Loops of side 160m were required to achieve this penetration, productivity would be about 20 sites per day. However **Fitterman et al (1991)** also reports the failure of TEM (and TCM sounding) in desert alluvium where the transition from partial- to full saturation is too gradual.

Two frequency helicopter borne EM surveys have been commissioned successfully to rapidly map the water table depth (between 25m and 100m subsurface) over large areas of western USA **(Paterson and Bosschart (1987))**. It is likely that the high costs involved can rarely be justified.

Strong GPR reflections are returned from the water table in sandy aquifers at shallow depths (less than about 15m); the technique is less successful in finer grained sediments **(Pullan et al (1994))**. Productivity is good, about 20km/day. GPR equipment is moderately priced (c £30 000) or can be rented at about £1 000/week.

Theoretically the EKS technique can determine the depth to the water table. However, recent work suggests that the characteristic EK signal can also be generated in partially saturated sediments above the water table **(Beamish and Peart (1996) and Peart et al (1996))**.

Seismic refraction is usually successful in determining the depth to a shallow water table (less than 15m or so), there being typically a velocity contrast between dry- and saturated alluvium of about 4:1 **(Haeni (1986) and Haeni (1995))**. The saturated section should be at least as thick as the unsaturated. The technique has been successful in deeper applications (100m reported by **Tucci and Pool (1985)**) but these require extensive geophone spreads (say 10 times the depth to be investigated) and large charges of explosive with the attendant problems of permitting, security etc. Seismic refraction is a relatively slow, labour intensive and hence expensive technique. It should be considered only when other methods appear unsuitable.

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6. SUMMARIES OF SUGGESTED PROCEDURES AND KEY REFERENCES

6.4 DETERMINATION OF GROUNDWATER QUALITY

Lateral extent of saline groundwater

TCM traversing will rapidly outline high conductivity zones indicating saline invasion to a depth of about 50m. The Geonics EM31 (single operator) equipment penetrates to 5m and offers productivity of about 10km/day and can also record a continuous profile. The Geonics EM34 (two man equipment) penetrates to 50m maximum with typical productivities of:

1.5km/day (observations at 25m, using all 6 coil separations/orientations, allowing crude 1-D modelling of conductivity variations with depth) or,

5km/day (observations at 25m using 2 coil separations only, allowing qualitative estimate of depth extent).

Alternatively TCM observations (employing all six combinations of coil separations and orientations) can be made at isolated observation points. **Van Overmeeren (1989 a)** obtained coverage at a station density of $1/km^2$ over $800km^2$ (at the rate of 40 stations per day). His results (contoured apparent resistivity maps for each orientation/separation) clearly defined the lateral extent of saline invasion and potable aquifers while the differences between contour maps indicated their depth extent qualitatively (subsequently confirmed by VES).

Barker (1980) recommends using only vertical coil configuration of the EM34 in this application (as this is less sensitive to coil misalignment and the assumption of low induction numbers is more valid in high conductivity zones).

DC resistivity traversing could be applied but the required galvanic contact with the ground slows progress (to about 1km/day) and may present problems due to high contact resistance in dry zones.

At a regional scale the lateral delineation of saline invasion can be determined by distributed VES/TEM, but these techniques are usually reserved for mapping the resistivity/depth distribution and for determining the depth of the salt/fresh interface.

ABEM surveys have been used to rapidly differentiate saline/brackish/potable zones in extensive shallow coastal aquifers **(Patterson and Bosschart (1987) and Oteri (1991))**

Vertical distribution of potable and saline groundwater and depth of interface

Traditionally VES have been applied, more or less successfully, to these problems **(Zhody et al (1974), Al-Ruwaih and Ali (1986) and El-Waheidi (1992))**. Equivalence is partially overcome by calibration of VES interpretation at isolated boreholes and then extension to remote soundings (assuming fixed resistivity). A common problem is suppression of the layer of intermediate resistivity (the aquifer) on the steeply descending curves characteristic of areas with dry/coarse overburden. The VES technique is most successful in shallow investigations (less than 40m or so). About 10 VES (with current electrode separation to 500m) can be made by 3+ men per day.

Occasionally IP soundings/traverses are made concurrently with VES to differentiate between low resistivity values due to saline water (low chargeability) and those due to clay (high chargeability) **(Roy and Elliot (1980))**. However, the procedure is slow (4 combined soundings per day) and the IP equipment is usually not readily available.

TEM offers far greater resolution of the vertical distribution of conductivity and is logistically superior to VES; productivity will be about 20 soundings per day (3 man crew). Depths of 150m may be investigated using a square loop of side 200m. **Fitterman (1987)** describes the constrained inversion of TEM data using isolated borehole information while **Fitterman and Stewart (1986)** undertook numerical modelling of TEM data to investigate the limits of detectability of a saline layer (in terms of both thickness and depth).

Software exists for the joint inversion of VES, IP and TEM soundings and such constrained interpretation will almost certainly yield reliable data on the resistivity distribution with depth.

<u>Degree of salinity</u>

The degree of salinity is indicated qualitatively by the aquifer's specific resistivity/conductivity. **Van Overmeeren (1989 b)** observed four distinct hydrogeological provinces in an unconsolidated coastal plain aquifer on the basis of **apparent** resistivities defined by VES: (a) saline (<1ohm.m) (b) brackish (3ohm.m) (c) potable (10ohm.m) and (d) shallow basement. **Gondwe (1991)** observed a correlation on the basis of **specific** resistivity (ie derived from interpreted VES): (a) potable water (30ohm.m to 80ohm.m) and (b) saline/brackish (<10ohm.m).

Oatfield and Czarnecki (1991) demonstrated strong inverse correlation between mapped transverse resistance (uniquely derived from VES curves) and sodium concentration (from borehole samples). The aquifer presumably showed little textural/lithological variation. **Al-Ruwaih and Ali (1986)** found strong inverse correlation between mapped aquifer resistivity (VES) and iso-salinity (based on borehole samples) and sited successful boreholes on the basis of transverse resistance maps (avoiding "lows" which implied thin aquifer, high clay content or saline water or a combination of these possibilities).

Van Overmeeren (1989 a) noted an empirical relationship between TCM-mapped resistivities and salinity conditions: (a) >25ohm.m implied freshwater aquifers more than 10m thick and (b) < 10ohm.m indicated saline water near surface.

Where certain parameters are known (aquifer porosity, clay content) quantitative estimates of salinity are feasible. **Hoekstra and Blohm (1990)** calibrated TEM data at several boreholes where the water quality was known and were able to equate observed specific resistivities with salt content (eg 8ohm.m represented a TDS of 500mg/L).

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6. SUMMARIES OF SUGGESTED PROCEDURES AND KEY REFERENCES

6.5 DETERMINATION OF AQUIFER PROPERTIES

It is stressed that aquifer properties derived from geophysical parameters will not be precise but they are probably adequate for useful approximations etc. The main advantages of such determinations are that they reflect bulk properties (compared with determinations made at borehole sites or on small core samples) and these aquifer properties are determined rapidly as a bi-product of geophysical survey.

Porosity

Using seismic velocity (eg the average velocity for an aquifer as determined by numerous refraction spreads): convert compressional (P-) wave velocity to bulk density using empirical **Nafe and Drake (1963)** relationship, then convert density to porosity by the formula **(Freeze and Cherry, (1979))** :

$$\Phi = 1 - (\rho b / \rho g)$$

where Φ is porosity and ρb and ρg are respectively bulk density and grain density (eg quartz for a sand aquifer). A systematic discrepancy between porosity derived thus and by standard hydrogeological methods probably reflects variable content of clay/silt and can be corrected for each new environment **(Ayers (1989 a))**.

Using formation resistivity (eg as derived from VES interpretation) : from the **Archie (1942)** formula modified for unconsolidated rocks:

$$F = 1/\Phi^{1.3} = \rho^{b}/\rho^{w}$$

where F is the formation factor, ϕ is the porosity and ρb and ρg are bulk formation resistivity (determined from VES interpretations) and pore water resistivity (determined from local well samples) respectively. **(Ayers (1989 a) and Ayers (1989 b))**.

Permeability or Hydraulic conductivity

Numerous direct (and rarely indirect) relationships between hydraulic conductivity and aquifer resistivity have been observed in granular aquifers. The precise nature of this correlation is governed by the average grain size and degree of sorting of the aquifer and hence will be site specific. It is also required that the electrical conductance of the groundwater remains constant or that an allowance (normalisation) is made for the variability of this parameter. **(Heigold et al (1979), Van Zijl et al (1981) and Allessandrello and Lemoine (1983))**.

Mazac et al (1990) investigated saturated sandy gravels and demonstrated a direct relationship of the form:

$$K\ (10^{-5}m/s) = 97.5^{-1} * \rho^{1.95}$$

where K is hydraulic conductivity and ρ is aquifer resistivity. Such an equation is site specific and the precise correlation between the two parameters should be determined anew in different environments.

Strong, direct (but usually non-linear) relationships between transverse resistance (thickness*resistivity product) and hydraulic transmissivity (permeability*thickness product) have also been demonstrated **(Ponzini et al (1984), Razack and Sinan (1988) and Niwas and Singhal (1985))**. This is particularly useful since transverse resistance may be obtained from VES data without knowledge of aquifer thickness. The transverse resistance values must first be normalised by dividing by porewater resistivity. Again, the specific relationships found will probably not be valid in different depositional environments.

EKS indicates permeability variations with depth on the basis of the rise time of induced EK voltages. At present such values will likely be absolutely correct only in the one hydrogeological environment for which the variables involved in the calculation (bulk moduli, shear modulus etc) are valid. Elsewhere the indicated permeabilities should be considered as relative determinations. Lateral permeability variations of any one horizon may be investigated by a traverse of EK observations, typically at an interval of between 15m and 30m. The technique is rapid (some 20 sites can be completed in one day) but our experience in unconsolidated aquifers suggests a penetration limit (using sledge hammer and steel plate) of about 40m. **(Beamish and Peart (1996) and Peart, Davies and Beamish (1995))**.

Aquifer grain size

Again this determination is primarily based on empirical observation of aquifer resistivity/conductance as observed with surface measurements. Provided the specific electrical conductance of the contained pore water remains relatively constant, then the bulk resistivity of the aquifer is representative of the aquifer's grain size (coarse grained material being more resistive than fine grained). The EM techniques (TCM and TEM) are more sensitive to subtle changes in electrical properties **(Haeni (1995))**.

Groundwater volume

An accurate residual gravity map of an entire basin is required (ie the Bouguer anomaly map minus the effect of regional and deep-seated structures); this will reflect the mass deficiency over the basin. Applying Gauss's law and incorporating values for the average density and porosity of basin fill and the average storage coefficient (either estimated or obtained from boreholes) the groundwater volume may be estimated **(West and Sumner (1972))**.

Aquifer storage change and specific yield

Temporal gravity observations (ie observations repeated at regular time intervals) at both bedrock- (reference) and aquifer stations reflect fluctuations in the water table (ie large variations are seen at the aquifer stations and much smaller variations at the reference stations when large scale movements of the water table occur) **(Pool and Eychaner (1995))**. It should be noted that extremely sensitive (and hence expensive) gravity metres are required for this type of survey. Also the technique is experimental, although Pool and Eychaner (op cit) claim they derived values of aquifer storage and specific yield similar to those obtained with aquifer test analyses.

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6.6 DETERMINATION OF THE EXTENT AND DEVELOPMENT OF CONTAMINATION

Groundwater conductivity is strongly affected (either increased or decreased) by the presence of even minor amounts of organic/inorganic contaminants, hence the exclusive use of electrical techniques in these applications.

Lateral extent of contamination:

TCM is most efficient for outlining lateral extent of conductivity anomalies:
Geonics EM31 (single operator) penetrates to 5m, rate of progress 10km/day, can record continuous profile. Geonics EM34 (two man) penetrates to 50m, rates of progress:

1.5km/day (observations at 25m, using all 6 coil separations/orientations, allowing crude 1-D modelling of conductivity variations with depth) or,

5km/day (observations at 25m using 2 coil separations only, allowing qualitative estimate of depth extent). **(McNeill (1989), Goldstein et al (1990))**

DC resistivity traversing requires galvanic contact with the ground and is therefore much slower than TCM; also possible problems with contact resistance in dry conditions. May be preferable to TCM where near surface is highly conductive (clays, saline soils) and may better resolve resistive bodies. Requires initial distributed sample VESs to determine optimum (focused) electrode separation. Rate of progress (3+ man crew) 1.5km/day (observations at 25m, two electrode separations). Rapid loss of lateral resolution as depth of investigation exceeds about 40m. **(Urish (1983)**

Where suitable borehole available (ie one that penetrates contaminated horizon) can apply screening-body technique (requiring placement of an electrode in borehole) and map potential distribution; improved lateral resolution of pollutant claimed **(Mazac et al (1989))**

Vertical extent of contamination:

TEM soundings are well able to resolve thin conductive layers down to 150m; rate of coverage about 25 soundings/day. VES more suitable for detecting resistive layers but the technique is "shortsighted" (ie a layer must be at least as thick as one tenth of its depth to be recognised). VES production rate is about 10 sites/day to current electrode separation of 300m. Combination of VES/TEM (or multi TCM observations) able to resolve equivalence problems through joint inversion.

Vulnerability of shallow aquifers:

Shallow protective clay layers rapidly outlined by TCM; their vertical extent determined by 1-D modelling of multi-observation TCM or complementary TEM/VES. Conversely, shallow sandy developments (zones of rapid recharge and hence vulnerable) are outlined as resistive features using the same techniques. Use of vehicle-towed multi-electrode DC resistivity techniques speeds survey (15km/day, full coverage with three electrode separations at 1m observation interval) but requires special soil conditions and equipment **(Auken et al (1994), Sorensen (1994), Christensen and Sorensen (1994))**

<u>Monitoring development:</u>

Ideally by regularly repeated (daily where critical) microprocessor controlled DC resistivity or equipotential observations across fixed grid of strategically placed electrodes. Alternatively can make repeated observations with rapid TCM methods. **(Osiensky and Donaldson (1994))**

<u>Miscellaneous comments:</u>

Commonly experienced is the problem of differentiating "signal" (response to contaminants) from "noise" (response reflecting the natural variability of clay and water content etc of the aquifer). In these cases the plume-like appearance of the anomaly (displaying progressive areal expansion and coincident weakening of anomalous values extending away from the source) has been shown to be a powerful discriminator. **(Greenhouse et al (1989))**.

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