Groundwater

Integration of geophysics and hydrogeology in the solution of regional groundwater problems

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Abstract. In the past, a tendency prevailed to carry out geological and geophysical investigations for groundwater independently of each other, or to regard the two methods as capable of replacing each other. This was partly because of the limited scope of many groundwater projects, and the preponderant emphasis on the element of prospecting. The present world-wide increase in the demand for groundwater is responsible for the growing number of groundwater exploration and development projects of regional size, requiring accuracy in the evaluation of available resources. For this purpose, an integrated application of geological, geophysical and hydrological methods and test drilling is needed. Such an integrated application was brought about, for example, in Israel, by the transition of groundwater exploration from the stage of providing solutions for local water supply problems to the stage of planning for regional schemes based on the utilization of groundwater.

In such projects, geophysics will be employed mainly, in co-ordination with geological studies and drilling, to establish a model of the subsurface structure of the region and the existing aquifers. It will also serve as an instrument for the solution of a variety of special hydrological problems, such as groundwater salinity distribution and localization of regions of groundwater flow.

It appears possible to develop a broadly standardized approach for the investigation of regional groundwater problems, grouping them into prototypes on the basis of similarity in their geographical-geological environment. Such standardization will also include geophysical procedure. It is felt that further standardization of geophysical field procedure, instrumentation, interpretation and presentation of results for those methods most frequently applied in groundwater investigations could make an important contribution to future large-scale use of geophysics in regional groundwater projects.

Résumé. Il y a eu tendance par le passé à entreprendre les recherches sur les eaux souterraines de façon indépendante du point de vue géologique et géophysique ou de considérer les deux disciplines comme interchangeables. C'était dû en partie à la faible envergure de nombreuses études des eaux souterraines et à l'importance exagérée donnée à la prospection. La demande croissante des eaux souterraines se traduit par un nombre de plus en plus grand de travaux d'exploration et de mise en valeur entrepris à l'échelle régionale, ce qui exige un haut degré de précision dans l'appréciation des ressources disponibles. A cette fin, on a besoin de méthodes fondées sur l'application intégrée des procédés géologiques, géophysiques et hydrologiques et de forages d'exploration. On s'est servi de méthodes semblables par exemple en Israel lorsque dans la recherche des eaux souterraines on a relégué au second plan les questions d'alimentation locale pour intensifier la planification régionale fondée sur l'utilisation des eaux souterraines.

Dans ces entreprises, la géophysique sert surtout à établir concurrement avec les études géologiques et les travaux de forage, une maquette de la structure souterraine de la région et des nappes aquifères existantes. Elle servira aussi à surmonter certaines difficultés spéciales d'ordre hydrologique comme la répartition de la salinité des eaux souterraines et la localisation des centres d'écoulement.

Il semble possible de mettre au point de puissants moyens de recherche pour trouver une solution aux problèmes des eaux souterraines à l'échelle régionale, en les groupant en prototypes en se fondant sur leur ressemblance dans leur milieu géographique et géologique. Cette normalisation comprendra aussi les procédés géophysiques. On croit qu'une normalisation plus poussée des procédés géophysiques, de l'utilisation des instruments, de l'interprétation et de la présentation des résultats obtenus par les méthodes les plus fréquemment utilisées pourrait être une contribution important à toute utilisation des procédés géophysiques pour la mise en valeur des eaux souterraines à l'échelle régionale.

A quarter of a century ago, when I first came into contact with problems of groundwater development in Israel, I became soon aware of a keen spirit of rivalry between geologists and geophysicists. Each side claimed to be in possession of the only effective means for locating groundwater resources and denied that the other could make a valuable contribution to the solution of the task at hand. Endless arguments were carried on both over past case histories and over future projects, with each side trying to score 'groundwater discoveries' without resorting to the co-operation of the other.

Before long I found that this situation was by no means unique in Israel, but prevailed in many countries engaged in the development of their groundwater resources. At that time, the science of hydrogeology was in its infancy and carried in many of its branches still much of the intuitive approach, if not to mention – alas – some deep-seated or subconscious involvement with water divining. While most geologists rejected this practice as unscientific and useless, many of them were overshooting the target in throwing geophysics overboard, too, as some kind of water divining with a pseudoscientific coating. They maintained that most geophysicists were hardly more than electricians, lacking adequate understanding of the earth's structure, or having, at best, only a rudimentary knowledge of geology, leading them to the use of geological technical jargon without ability to appreciate its meaning.

Wherever the opinion of such geologists gained the upper hand, the result was frequently indulgence in expensive and excessive test drilling in areas where a geophysical investigation could have easily solved the problem of subsurface structure and provided important clues prior to drilling.

Geophysicists, on the other hand, regarding disparagingly the limitations inherent in geological methods, were frequently inclined towards a look-for-the-needle-in-the-haystack approach, promoting extensive and expensive surveys with vociferous salesmanship, often superior to that practised by their geological competitors.

The planning engineer, confronted with such a situation, found himself in a serious predicament when he had to decide on the approach to groundwater surveys. In the first place, he tended to be. and in many cases still is today, highly suspicious of the whole subject of groundwater. Because of his engineering education he was frequently heavily biased in favour of

development of surface water, a resource exposed to view and direct observation, and lending itself to handling and management through construction of solid engineering and hydraulic works. Groundwater, on the other hand, being an invisible resource demanding an indirect approach and more sophisticated method of evaluation, was regarded by many engineers as having some mystical qualities. Groundwater hydrology, a subject which could appeal to the engineer through its use of physical principles and its quantitative approach, was still beginning, and few geologists or geophysicists were in a position to speak its language. Therefore it is not surprising that engineers often took an unbelieving attitude towards any forecasts on groundwater occurrence and towards handling of groundwater as a phenomenon of natural science.

Wherever conditions were such that groundwater nevertheless had to be prospected for because surface water was not available, the engineer's decision on a means of approach was subject to conflicting considerations. On the one hand, the engineer tended to be sceptical about the results of geological investigations, achieved through the use of nothing but hammer, lens, barometer and maps, and was mostly inclined to rely more on the explanations of the geophysicist, working with more complicated and impressive apparatus and equipment. On the other hand, he tended to be reluctant to allocate the funds required for geophysical surveys, which were in his opinion out of proportion to the benefits to be expected from them.

One of the major reasons for this either-or attitude in the choice of prospecting methods was the limited magnitude of the problems to be studied, both in terms of terrain to be covered, and of water required. Even today it seems justified to rely on geophysical methods alone for investigation of a small area in a large plain, while in other terrains, areas with good outcrops may lend themselves to rather complete subsurface analysis through geological methods only.

Today, it is realized in most countries that groundwater exploration, and in particular, groundwater exploration on a large, regional scale, requires the co-operative and co-ordinated efforts of geologists, geophysicists and hydrologists, under the general guidance of the planning engineer. While the regional size of groundwater projects has been the prime factor in bringing about the integrated application of the disciplines involved, progress has been made in geophysical methods and instrumentation, in the specialization of the geologist in the application of geological considerations to groundwater problems and in the coming of age of groundwater hydrology as an instrument for the quantitively exact planning of the utilization of groundwater resources. The integrated application of all three methods is gaining ground with the general trend towards team work in modern science and technology, and the spirit of co-operation is superseding now the atavistic instinct of individual treasure hunting.

Nature of regional groundwater projects

After this introductory digression into the not too remote past, let us now have a closer look at present-day groundwater problems and examine the ways for their solution by the integration of geological, hydrological and geophysical methods. Regional, or large scale, groundwater problems are studied in our times at an ever increasing rate by central and provincial administrations, large co-operatives and by agencies for international co-operation ranging from the United Nations and its specialized agencies to regional organizations of states and to technical co-operative plans based on bilateral agreements. This increase in interest in large-scale groundwater programs has certain advantages over surface water projects. While the benefits of surface water programs are attained only after the completion of large and expensive construction works demanding heavy capital investments, groundwater planning can be developed gradually, with relatively low capital investment, and benefits are obtained even from the first exploratory stages. Thus, increase in available water can keep pace with the necessarily gradual advances in methods of use and application of water. Moreover, groundwater planning is characterized by greater flexibility and built-in safety factors against periodic droughts.

There are also a number of significant differences between local and large-scale groundwater development programs:

1. Large-scale groundwater development programs are more ambitious, aiming not only at supplying water but also at very fundamental economic and social transformations.

2. Large-scale groundwater programs will affect in the course of time the entire hydrological regime of the region, including its surface water resources, and require therefore a full assessment and forecast of the effects of exploitation on the water resources in terms of quantity and quality of available water in various stages of the project.

3. The variety of conditions encountered on a regional scale calls for an integrated application of geological, geophysical and hydrological methods, and no single method will provide the desired solution of the problem at hand.

4. A faulty appreciation of a local groundwater problem may result in economic losses, but mistakes in the regional appreciation of groundwater resources may well be disastrous.

Regional groundwater problems exist and await solution today in many different parts of the world. Though many of them occur in the dry zones, they are not limited to them. They differ widely in their purpose and nature, encompassing such diverse projects as the supply of natural hot groundwater for heating of homes and hothouses in Iceland, the finding of adequate water supplies for mineral processing in the Atacama Desert of Chile, the evaluation of groundwater supply potential for irrigation in sparsely inhabited and underdeveloped lands, the replacement of polluted and unreliable surface water supplies by pure and easily managed groundwater for municipal supplies in tropical, subtropical and temperate belts, and studies to prevent water logging in intensively irrigated areas in river basins. Thus, regional groundwater projects may vary considerably in their general objectives and in their geographical-hydrogeological environment.

The approach to regional groundwater projects

Neglecting for this paper the economics of groundwater supply, the objective of a regional groundwater program is to make the desired amount of groundwater of suitable quality available where and when required. This purpose will be achieved through the construction of the means for withdrawal of water from the ground, which will in most cases be wells of various types. This groundwater development phase must be based on an investigation of the groundwater possibilities of the region to find out whether the desired amount of groundwater will be available and

where it can be most conveniently exploited. This is the groundwater survey phase of the project. The survey phase can be carried out prior to the development phase, or simultaneously with the earlier stages of the development phase, or preceding it but with a considerable amount of overlap. The approach to be selected will depend on such factors as the amount of data available at the beginning of the project, the urgency of supplying water, and the system for financing the project.

The consecutive approach will be preferable where relatively few data on subsurface structure and hydrology are available, where the urgency to improve water supplies is not very great, and where funds for pre-investment studies and for actual development will come from different sources and will be made available at different times. The simultaneous approach should be selected where a considerable amount of previous data permits the immediate initiation of a development drilling program, where there is great urgency to increase available supplies immediately and the hydrological situation supports such increase, and where funds for both investigation and development can be made available at the same time, or derived from the same source.

The advantages of the consecutive approach are that the results of a fairly complete investigation, including definite recommendations as to the most convenient zones for groundwater exploitation, and a groundwater balance with estimates for exploitable quantities will be available prior to the initiation of a large drilling program. The disadvantage is, however, that such an investigation program will not normally provide for an adequate amount of drilling of large-diameter wells, and therefore the results will remain largely untested, or based on partial conclusions from small-diameter exploration holes.

The simultaneous approach takes advantage of the possibility to step up water supply gradually and concurrently with other advancements in the development of the region. It calls for continuous and constant mutual adjustment of the investigation and drilling program through feed-back of forthcoming results. More sophistication and a higher amount of flexibility in the management are required, but results will be more conclusive and more adequately tested.

Integration of geophysics into regional groundwater projects

All regional groundwater projects will require at various stages the service of geophysics. It will be used mainly as an aid in the determination of subsurface structure, of the horizontal and vertical distribution of aquifers and their regional boundaries and for the location of drilling sites. It will also assist in the solution of a multitude of special problems, such as tracing fresh and saline waters in aquifers, analysis of the mechanism of springs, location of regions of groundwater outflow into the sea or lakes, detailed investigation of areas for artificial groundwater recharge, and as an aid in well construction.

A wide range of geophysical methods is available today for application in groundwater exploration, and these will be dealt with in detail by other papers presented to this Conference. Surface electrical resistivity methods are still the most widely used for general and detailed subsurface reconnaissance because of the low expenses involved and the wide range of applicability. Important advances have been made lately in improving the use of such methods to reach greater depths. Relatively inexpensive seismic refraction methods are now employed at an increasing rate, alone or in combination with resistivity methods. Magnetometric methods might be useful in a number of special cases. Electromagnetic (AFMAG) methods have been recently introduced for locating fracture zones. Gravimetric methods could be useful for a structural analysis of the deep-seated base and boundaries of aquifers overlying, for example, crystalline rocks. Thermometric methods have been applied recently with success in hydrogeological investigations, and radiometric methods can certainly contribute to the solution of some specific problems. Logging methods taken over from the petroleum industry have made an important contribution mainly by permitting the use of the more economic rotary method of drilling for exploration purposes, especially where considerable depth must be attained. One of their important assets is the possibility to determine formation fluid salinity.

Geophysical investigations are an integral part of regional groundwater surveys, and must be carried out in close and continuous co-ordination with hydrogeological studies and drilling. Although this statement may appear commonplace to this audience, it cannot be omitted. There are still many cases where geophysical work is carried out completely detached and isolated from other components of groundwater development programs, and instead of making reliable additional data available, the results only add confusion because they consist of ambiguous data out of context with the focal problems of the project.

In most cases, a groundwater project will start with a preliminary assessment of hydrogeological conditions as known from pre-existing surface and subsurface data. This will lead to the definition of the additional geological and hydrological information required for the preparation of a drilling program, a groundwater balance, and a practical long-range plan for groundwater utilization and management.

The means to obtain this information will then have to be determined. These will include geological and geophysical studies and multipurpose test drilling. Geological methods will be mostly of the subsurface type, characterized by the need to arrive at conclusions from extrapolation of surface observations, and interpolation between occasional subsurface data. Geophysical methods will be called upon to prove or disprove working hypotheses arrived at by geological reasoning. As geophysical methods are indirect by nature and subject to interpretation, only the closest contact between the geologist and the geophysicist can ensure optimum results.

It is desirable to include at an early stage geophysical measurements that can be correlated with reliable subsurface data from well-documented boreholes. In case no such boreholes exist, or they do not provide information to sufficiently great depth, a number of test holes should be drilled immediately. The sites for these boreholes should be selected so that the most information can be obtained besides the geological-geophysical correlation. They should permit the observation of water levels in various aquifers, sampling of water for quality, and they should be drilled with a diameter that permits the performance of pumping tests for the determination of aquifer transmissivity and storativity. Finally, it is always desirable that an exploration well be located so that it can later be used as a production well. It will not always be possible to attain all these objectives in a single well, and clear priorities must be established when the location is defined and the drilling method is selected.

On the basis of these correlation studies, methods may have to be modified, or additional methods may have to be introduced. After completion of this methodological stage, the first phase of the regional geophysical survey can begin. The purpose will be a rapid general reconnaissance of the subsurface, laid out as a network or a series of profiles following important structural or hydrogeological trends. Although the objective of this phase is a general one, objectives should be clearly defined, and it might be useful to include some more detailed work in the form of a more closely spaced investigation in areas assumed to be of special interest in connection with the groundwater hydrological regime, for example sections along which significant groundwater inflow or outflow is to be expected.

The plan for this first stage of geophysical work should be prepared in co-operation between the hydrogeologist and the geophysicist, keeping in mind the objectives. It will in most cases be founded on some working hypothesis of an aquifer model, involving many assumptions, which investigations will have to prove or disprove. Blind adherence to this plan should be avoided, and it should be adjusted on the basis of a continuing evaluation of incoming results from geophysical and other sources. Thus, the plan may be modified to investigate unexpected anomalies further and the test drilling program will have to be adjusted accordingly. Similarly, the geophysical work plan will have to be revised when the progress of test drilling produces additional data for correlation.

The results of this reconnaissance stage of geological, geophysical and drilling data should be synthesized to provide a more advanced aquifer model that could serve both as a basis for the preparation of an ample production-well drilling campaign, and for a regional groundwater resources assessment, including a first hydrogeological balance.

A second major phase of geophysical work will follow after the necessary assessment and changes in the investigations are made in connection with the groundwater exploitation plan. Projected areas will be investigated in detail by the suitable selection of drilling sites to optimize conditions for water supplies in relation to areas of demand. At the same time, emerging specific hydrological problems, such as the position of fresh water-salt water interfaces, may require further work.

With the completion of this second phase, there will be a reduction in geophysical work, and emphasis will pass to the application of logging methods as an aid in well construction and subsurface correlation. Radioactive tracers may be applied for the study of the hydrological regime. At a still more advanced stage, when groundwater resources utilization reaches its optimum, very detailed prospecting of limited areas designed for artificial recharge may be required.

Integration of geophysics into groundwater development in Israel

The important part played by geophysics in the development of Israel's water resources is explained by the fact that these consist mainly of groundwater since, of an estimated total annual water consumption of 1.45×10^9 cubic meters, about 80 per cent is derived from groundwater sources, including spring flow. Out of the necessity to rely mostly on groundwater resources for our water supply, we in Israel have learned the many advantages groundwater presents over surface water, and are today even converting surface water into groundwater, for storage and upgrading of quality.

Prior to independence, demands were rather limited, and water had to be supplied as close as possible to the area of demand. The main problem was the location of successful boreholes on the land held by a certain village. This period, between the early nineteen-twenties to the late nineteen-forties, became the period of 'groundwater discoveries', during which geologists and engineers became familiar with the major prospective aquifers. Geophysical methods, exclusively surface resistivity were applied for the location of boreholes, but lack of co-ordination with geological studies and lack of sophistication in the geological reasoning itself, resulted in some serious setbacks.

In the decade following independence in 1948, it became possible to approach water resources development on a regional and even national scale. Activities started with the planning and execution of regional schemes based on groundwater, spring flow and in some cases on limited use of surface runoff. These schemes were designed to fit at a later date into a national master scheme based on the integration of the regional schemes with the addition of part of the flow of the Jordan River. This national water grid, which today is in a very advanced stage of execution, makes full use of the long- and medium-term storage facilities provided by underground water reservoirs. This is a feature of foremost importance in a semiarid country where a sequence of years with below-normal rainfall is quite common. Using the aquifers as conduits, water can be withdrawn in convenient areas close to the points of demand. The natural cyclical flow rate of surface runoff and springs resulting from alternating wet and dry seasons is transformed through appropriate aquifer management so that losses of water are minimized and supplies are made available not only where, but also when required. Groundwater appearing in the form of brackish springs has been upgraded by interception through wells ahead of areas of mineral pick-up, or through redistribution of excess salinity by surface and underground mixing.

The creation of a national water planning agency has greatly contributed to the integration and full co-ordination of geological, geophysical and hydrological work in connection with the planning and execution of this national water scheme. Geophysical methods, especially the surface electrical resistivity methods, have been extensively used during various stages of planning and execution.

The two most important aquifers in Israel are the sand-sandstone aquifer of the coastal plain, and the limestone-dolomite aquifer of the hill region. The coastal plain aquifer consists of a sequence of clays, loams, sands and calcarenites of Pliocene-Pleistocene age, underlain by impervious Miocene-Pliocene clays and sandy clays. Its thickness reaches 150 m near the seashore, and it wedges out gradually towards the foothills in the east. At the same time, the percentage of marine calcarenites in the total section decreases from west to east. The aquifer is subdivided into various subunits by impervious layers. During the reconnaissance stage, resistivity methods were applied to survey those parts of the plain where few wells had been drilled, or where existing wells had not penetrated the full section of the aquifer. The resulting data were used in the preparation of drilling program for production wells. After stepping up exploitation, when water levels started to respond, a network of observation wells was drilled to facilitate full control of the underground water flow. At this stage again wide use was made of surface resistivity methods to locate well sites, and of logging methods for correlation. At the

same time a full investigation of the position of the fresh water - salt water interface along the seashore was carried out, supported by ample use of surface resistivity measurements, that could already rely, in this case, on a fairly complete picture of the lithostratigraphic character of the subsurface. Very detailed geophysical investigations were carried out at a still later stage over areas where groundwater recharge operations are to be carried out. At the present time, the coastal plain aquifer is fully controlled and managed within the framework of the national water grid. Natural outflow to the sea is kept to the minimum necessary to prevent the penetration of the salt water interface beyond a predetermined limit. In the future, this outflow, too, will be utilized through a string of shallow wells constructed near the coastline. Seasonal floods and surplus from surface flow are recharged into the ground.

Geophysics took a still more important part in the development of the second major aquifer, the limestone-dolomite aquifer. It consists of several hundreds of meters of limestones and dolomites, interbedded with chalk and shale, of Upper Cretaceous age. Outcropping over extensive areas in the hill country, it plunges below the coastal plain in a series of anticlines and synclines, disappearing below an impervious cover of chalk and shales of Cretaceous to Tertiary age, overlain in turn by a thick fill of the coastal plain. In central and especially in northern Israel, this formation is also intensively faulted, together with all overlying pre-Pleistocene formations. This faulting is responsible for the formation of interior depressions of rift-valley structure, like the Emeq Yesreel and the Jordan Valley.

Suggestions that the dolomite-limestone formations of the hill country could be a major source of groundwater had already been voiced in the early nineteen-thirties. But many of the earlier tests failed because of lack of knowledge on the aquiferous properties of the various formations, insufficient data on their thickness, absence of information on the position of the regional water-table and sometimes because of overestimation of the importance of single factors believed to assure success of drilling, such as location of boreholes on fault zones, or drilling to the depth of the water-table without due regard to the formations encountered there. But some significant successes had already been obtained prior to independence, mainly in the Carmel and Emeq Yesreel areas and in the central Coastal Plain, near the mountain border.

With the initiation of the regional schemes, systematic investigations were undertaken to prepare the ground for large-scale exploration of this aquifer. Geological mapping and subsurface investigations based on all existing boreholes were directed towards a detailed structural and stratigraphic analysis that would permit reliable forecasts to be made on drilling locations in and near outcrop regions. A special micropaleontological study was made to subdivide the lithologically uniform chalk-shale complex overlying, with variable thickness, the limestone-dolomite aquifer, and to make depth-to-aquifer estimates more reliable. Last but not least, a good resistivity contrast between the aquiferous dolomite-limestone complex and the nonaquiferous overlying strata made it possible to trace this contact in the subsurface of the plain. Improvements of methods made it possible to locate it to depths around 1000 m.

Regional integrated studies of this type were carried out for Western Galilee, the eastern part of the Coastal Plain, the interior valleys of Galilee, the Jerusalem area and the northern Negev lowlands. As a result, zones of structural highs could be defined where the depth required for reaching the aquifer would be minimal. It was soon found that such zones were also those where the highest permeabilities were encountered, and where, in areas of salinity problems, salinities tended to be lowest. In many cases, however, it was found advantageous to drill at locations of greater aquifer depth to reduce surface conduit requirements by using the aquifer itself as a conduit.

Today, the regional development program incorporating construction of the appropriate number of wells is practically completed, and the aquifer is under a fully controlled administration. Its main functions are holdover storage from the rainy season into the season of peak demand, storage of surpluses from the national system for priority use in dry years, and salinity control. Spill from the few large springs that in the past constituted the principal outflow is minimized through proper management of pumping and, when it occurs, recycled into the aquifer through recharge wells.

Thus, the integrated application of geological, geophysical and hydrological methods, guided by an engineering approach not limiting itself to conventional solutions and boldly exploring the special characteristics and advantages of groundwater resources, has helped to establish in a country poor in surface water resources and visited by frequent droughts a water supply system which efficiently assures supplies even under short- or long-term emergencies.

Geophysics in typical regional groundwater schemes

Despite the variability of projects, and each groundwater problem's individuality, a great many of the regional problems occur in a rather limited number of geographical-geological settings. The reason is, of course, the necessity for coincidence of certain factors to make the regional groundwater problem arise: the region must be fairly inhabited or inhabitable, a plain or gentle relief, and in a climatic zone where water deficiency prevails at least during a part of the year. This rules out high latitudes, high mountain regions and areas with plentiful year-round rainfall. From a practical point of view, the most frequently encountered environments of regional groundwater problems can be listed as follows: (a) coastal plains, deltaic areas, large alluvial plains of giant river systems; (b) continental shield regions in temperate, subtropical and tropical belts; (c) regional sedimentary basins; (d) volcanic plateaus; (e) limestone terrains; and (f) oceanic islands.

Each one of these typical groundwater-problem environments presents a characteristic set of problems to the hydrologist, the geologist and the geophysicist. Let us. in conclusion, have a brief look at the geophysical problems to be solved.

In coastal plains, deltaic areas and large alluvial plains, the principal task will be the definition of the ultimate basis of the aquiferous system, and the distribution of aquiferous, non-aquiferous and intermediate sediments within it. The coastal plain of Israel quoted above might serve as an example. Geophysical methods applied will have to reach frequently to depths exceeding a hundred, and sometimes several hundreds of meters. Where the subsurface stratigraphy along the coast is sufficiently well known, it is possible to locate the fresh water - salt water interface by the resistivity method.

Continental shield areas are characterized by the presence of practically impervious crystalline rocks at relatively shallow depth, under a cover of various loose sediments. These may be

glacial deposits in temperate-cool zones, desert weathering products in the subtropics and laterites in the tropics. One of the questions to be solved by geophysics is again depth to bedrock. Geophysical methods should be able to pick up with great precision irregularities in bedrock topography, and the character of the overlying sediments. Furthermore, the nature of the uppermost portion of the bedrock, whether compact or decomposed and fractured, will also be of importance. Examples well known to this audience are the demands made on geophysics to analyze the nature of the glacial deposits in the northern part of the North American continent. In Africa, the thickness of the disintegrated upper zone of the crystalline bedrock underlying the laterite is of importance in many cases. While methods applied will not be required to reach to great depth, a very detailed picture is required in this type of situation to obtain useful groundwater quantities.

Regional sedimentary basins are, for example, those found in the bowls formed by the continental bucklers. Famous aquifers occurring in this type of setting are the continental sandstones of Northern and Central Africa (Grès Continental Intercalaire), the Dakota Sandstone of the United States midcontinent region and the aquifers of the great Australian artesian basin. Changes in thickness and lithofacies, depth to upper and lower contact and possible buried irregularities of the substratum forming barriers or subdivisions of the aquifer will be the main objectives of geophysical investigations, and these should be able to penetrate to considerable depth.

At present, resistivity methods are widely applied in various forms to resolve these types of problems, but seismic refraction techniques are becoming more and more accepted, especially for the problems presented by the continental shield areas.

In volcanic terrains, the questions to be answered will concern the subsurface distribution of aquiferous lavas and nonaquiferous lavas and pyroclastics, as well as the directional trends of dyke systems. The question of how much the 'popular' resistivity methods can resolve these problems does not seem to be clearly answered as yet. It might be useful to support or to replace the resistivity method by refraction and magnetometric methods, or gravity observations.

While the resistivity method has given some excellent results in the investigation of the limestone aquifers of Israel, there also seems to be still a large amount of inconclusive results from other areas, and one can find strongly divergent ways of interpretation in various case histories. Apparently there is no method yet that can distinguish between aquiferous and nonaquiferous zones in a hard calcareous rock. Methods that could point to major fissure zones promoting development of karstic channels would be able to make an important contribution.

Oceanic islands are mentioned as a separate type mainly because of the particular hydrological problems they present. From the point of view of geological-geophysical investigation they do not differ from continental areas of similar geological character. Most of them will be volcanic. Geophysics may be called upon to localize centres of groundwater outflow to the sea in addition to the usual subsurface analysis, and to determine the position of the fresh water - salt water interface.

It is hoped that the development of standardized, yet flexible, type approaches to regional groundwater problems will contribute to the successful conclusions of the many projects to be expected in the future. May a geologist express, in conclusion, his wish that geophysical instruments, field procedure and presentation of results be likewise standardized in a way that would permit easier regional and inter-regional interpretation and comparison of results, thereby promoting the widest possible use of geophysics in regional groundwater exploration.

Discussion on Integration of Geophysics and Hydrogeology in the Solution of Regional Groundwater Problems by Z.L. Shiftan

Utard, M., Geoterrex CGG, France. I would like to have some typical examples of a hydrological problem in Israel. I think there are many tricky problems and I should like to know more about the way you are solving them, even if not always by geophysical techniques.

Shiftan, Z.L. I should probably answer the question in completing a few details that I had to omit from the oral presentation and going a little bit into descriptive detail.

Israel is fortunate in having two good main aquifers; one is the sandstone aquifer of the coastal plain, the other is the limestone dolomite aquifer of the mountain region. The relation between these two aquifers is such that the limestone aquifer dips down in a series of flexures, anticlines and synclines, below the coastal plain with an intervening quite thick packet of impervious chalk and marl formations (see Figures 1 and 2).

I think one of the disadvantages for taking an Israeli example for the integration of geophysics into groundwater is that groundwater development in Israel has already passed its heyday and now the subject of proper application of various geophysical methods in regional groundwater exploration problems is arising. During the main stage of groundwater exploration and development in Israel, the only method used was the resistivity method. Very rarely, in very few cases, did we apply gravity methods for locating areas of various densities, some magnetometer surveys to help in pinpointing faults and some shallow seismic refraction.

One very important hydrogeological type problem was the depth of the limestone aquifer, below its impervious cover. It was essentially a two-layer problem. Although the upper conductive chalk-marl sequence may be very thick, it is possible to pick up, by using the resistivity method, the contact with the limestone very well down to a depth of about 1000 meters. This does not sound too astonishing anymore today, but it certainly did about ten to fifteen years ago when we ourselves could not believe that this would be possible. I think that this was the most important contribution of geophysics in the development of the limestone aquifer.

May I add a word at this time about the present situation of exploitation of the limestone aquifer. One of the most interesting things about groundwater development in Israel is that the exploitation of the aquifers has been pushed to the utmost possible limit. The limestone aquifer is naturally drained through springs, giving rise to short coastal rivers, or directly through outflow into the sea.

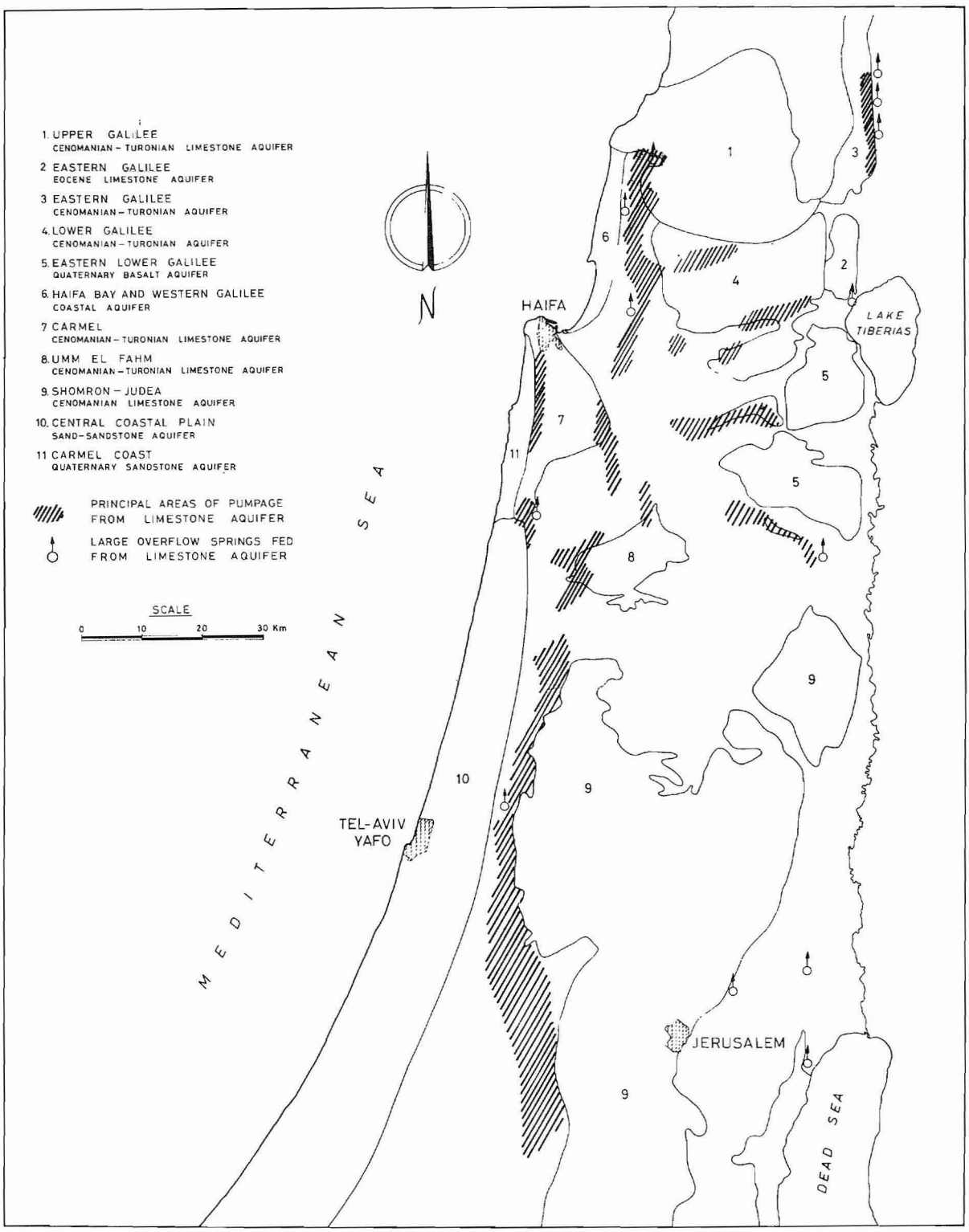


Figure 1. Israel – map of important aquifers.

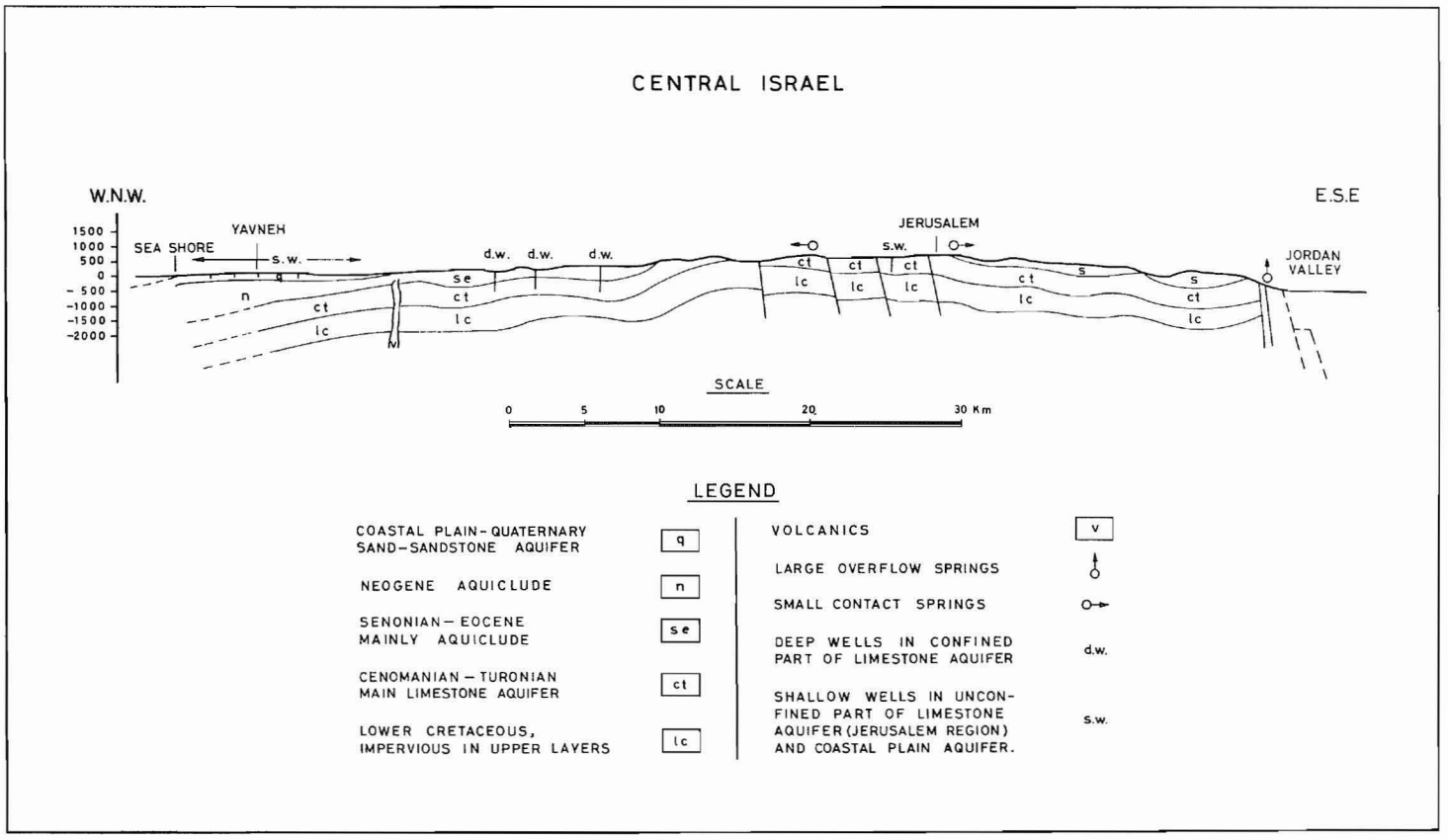


Figure 2. Hydrogeological cross section through Israel.

The latter is the rarer case, the more typical case is drainage through springs. In a semiarid climate with a pronounced rainy winter season and a long dry summer season, there is a cyclic variation in the discharge from the major springs, with the highest discharge occurring in the springtime when the soil is still wet from the winter rains and irrigation needs are at the minimum. On the other hand, when irrigation is very much needed, in summer to late fall, the spring discharge is at its lowest. The problem presented by this phase difference was solved by drilling wells into the limestone and introducing a pumping regime and managing the aquifer at a slightly lower level than the level of the springs. Water would be pumped out during the irrigation season to make room for the winter rains which fill up the reservoir again and bring it just up to the original spring level, thus preventing massive losses to the sea. Today, the pumping regime is calculated and planned in such a way as to optimize the use of the limestone aquifer as a medium-long-term storage reservoir, to prevent losses to the sea, and also to arrive at a desirable salinity, using it for subsurface mixing of very fresh and slightly brackish water.

The coastal plain aquifer has also been explored extensively with the resistivity method. This was a little more difficult, because it is composed of frequently alternating layers of sandstones, sands, loams and clays interbedded in a very irregular succession. The results obtained from the resistivity surveys was not always very satisfactory. Nevertheless, during the systematic investigation of the coastal plain for the drilling of a network of observation wells, with a 2-kilometer spacing to permit controlled management of the aquifer, the resistivity method was very successfully used to locate the observation wells properly and to assure that they penetrate the proper horizon.

The main hydrological problem of a coastal plain aquifer is very well known to all of you. It is the problem of the fresh water-sea water interface. The problem arises in a very acute manner when exploitation of the aquifer is to be driven to its utmost limits. New methods were looked for to solve the dilemma presented by the need, on one hand to maintain a certain outflow to the sea, in other words, to maintain a certain groundwater level above sea level in order to prevent salt water intrusion, and on the other hand, to prevent outflow of groundwater to the sea in order to assure maximum use of available groundwater. It will be necessary to maintain a certain groundwater gradient towards the coast to safeguard the aquifer against seawater intrusion. The question how this water can after all be made available for use was solved by an investigation supported by a United Nations Special Fund grant. The basic idea was to catch the outflow to the sea just short of reaching the seashore by a groundwater interceptor, which would collect this water and from which it could be pumped to nearby areas of demand or water distribution systems. It was assumed at the beginning that this coastal collector would have to take the form of a horizontal drainage installation but experiments showed very clearly that this is not necessary and that a line of wells would be just as efficient as a much

more expensive horizontal collector. This coastal collector is now in the state of construction.

May I mention one more point about aquifer management. I touched briefly in my talk upon the integration of surface and groundwater in total water resource management. An important point in the development plan for Israel's water resources was to take full advantage of the long-term storage facilities offered by existing groundwater reservoirs, disregarding for a certain time what is normally called 'safe yield'. The term 'safe yield' itself is very vaguely defined and it contains factors which have no physical meaning, being based on economic considerations and therefore are subject to changes in time. Taking advantage of the large volumes of water stored in the major groundwater reservoirs, these were drawn upon quite in excess of the amount of the annual replenishment. In this manner quantities of water were made available to users which otherwise large and costly surface water schemes would have provided at a much later date. In practice, this means that aquifers were overexploited consciously to provide the amount of water planned to be available on completion of the National Irrigation Master Plan, based on integrated use of surface and groundwater, including the use of that part of the Jordan River flow allocated to Israel. Thus, large amounts of water from groundwater sources became available many years prior to the completion of the surface water development schemes. The effect was a more convenient distribution in time of heavy capital investments and, at the same time, an immediate gain from the expansion of agricultural production. Today, these reserves exploited in advance of the completion of the surface water scheme are now replaced from the winter surplus water, and aquifers are replenished mostly through wells in order to build up again the stored reserves, to be used, say, during a future spell of dry years. This system provides, therefore, an important factor of safety against periodic droughts, or recurrent years of precipitation deficiency. In fact, a dry year – a very common occurrence in the Middle East – hardly affects Israeli agriculture at all.

Collett, L.S., Geological Survey of Canada. Is your groundwater movement controlled by some central system?

Shiftan, Z.L. This is also a point I should have mentioned right at the beginning. Water in Israel is defined by law as public property, not as state property, and is managed by the National Water Authority. All water resources are indeed managed centrally, and there is a nationwide water distribution system allowing for great flexibility in the absorption of water input and the meeting of demands.

Question. On the salt water - fresh water problem, just before the fresh water reaches the ocean do you pump the fresh water back into the aquifer upgrade?

Shiftan, Z.L. This water is returned into the general network. Whether it is used for irrigation or recharge upgrade will depend on the season.

Question. Now, you have your wells almost at the discharge point into the sea water. Haven't you aggravated your problem in that you may introduce sea water to those wells?

Shiftan, Z.L. Of course, this is a very delicate hydrological problem that must be very closely watched. I mentioned before a system of observation wells is maintained to check the salt water - fresh water interface constantly. The pumping regime is very closely controlled by these wells. You should also realize that the water-table does not change abruptly in terms of meters during a year, it is a matter of a meter a year. We notice small trends in groundwater movement, this is adjusted for in regulating the pumping regime.

Raza, S.H., Geological Survey of Pakistan. I would like to know the technique that you have been using for interpreting your resistivity curves.

Shiftan, Z.L. The geophysicists used, up to a very short time ago, the Wenner array and then master curves for the interpretation. Now we are introducing the method which Dr. Flathe has presented here, changing over to the Schlumberger-configuration and using the computer for better interpretation than by matching the curves.

Raza, S.H. What special techniques do you apply for overcoming the high stake resistances in arid regions?

Shiftan Z.L. I must again excuse myself, I cannot answer you as a geophysicist, only as a geologist. I know that, in many cases, porous ceramic pots with copper sulphate solution are used instead of metal electrodes and this works very well, thus eliminating polarization effects.

Question. What were the resistivities involved in the coastal plain area of Israel?

Shiftan, Z.L. The resistivity of the limestones varied between 15,000 to 25,000 ohm centimeters and resistivities of the overlying chalk and marl beds ranged between 2000 to 5000 ohm centimeters, a very good contrast.

Raza, S.H. Which interpretation techniques did you find most successful: curve matching based on theoretical work or empirical method based on curve inflections?

Shiftan, Z.L. Most of our experience is with the Wenner array, but our experiences over the past year using the Schlumberger array are very encouraging and showing better results. For interpretation, curve matching is the common technique, but is checked by actual correlation measurements wherever this is possible.

Raza, S.H. Was there good correlation with the geological information?

Shiftan, Z.L. Yes. I should mention one more possible application. In cases where we do know the subsurface stratigraphy very well and we have had borehole information, we can sometimes, over a neighbouring area, find salt water by surface resistivity. But this comes at the stage where we already know the subsurface stratigraphy, otherwise it is difficult to interpret.

Utard, M. I disagree. I think that it is not necessary to know the conditions very well to map the invasion of salt water, especially in muds.

Shiftan, Z.L. The solution is to set up measurements for a depth where you might expect salt water. If you have at this depth fresh-water saturated clays, or salt-water saturated sands, will there still be a difference?

Utard, M. Yes.

Shiftan, Z.L. In the coastal plain of Israel, the base of the aquifer is a very thick clay sequence, probably up to hundreds of meters. This may be the reason that it is sometimes difficult to distinguish clays from salt-water saturated sands above. In practice, however, the problem does not arise, as the subsurface conditions are always fairly well known.

Collett, L.S. I would like to ask Mr. Shiftan to expound on his ideas of standardization of geophysical methods for groundwater.

Shiftan, Z.L. While listening to interesting papers presented here on the use of geophysical methods for groundwater, the question arises: where do we groundwater hydrogeologists go from here and where is the place of geophysics in groundwater from now on? What worries me a little bit is that geophysics after all is business, much more so than, shall I say, geology, and business is in mining and in oil. Definitely groundwater business cannot compete with these two. I cannot blame the geophysical art for neglecting groundwater while there are these fantastic prospects of interesting and attractive work in the mining and oil fields. Nevertheless the services of geophysics are needed in groundwater exploration. These services, in many cases, take the form of a man who has bought himself a little Terrameter or a little Gish-Rooney apparatus and has put an ad into the paper to the effect that he is open for business. Without overstating the case, what mostly would be required is people from larger geophysical outfits who would do some thinking about the groundwater problem at hand and not just operate a little tool of the 'drugstore prescription' type. I think this is in many countries and cases the most serious problem. If we could assure the attention that the groundwater problem requires instead of getting a little, shall I say, 'medical practitioner' who operates a tool that he does not understand very much about, then we would be much better off. We must secure the cooperation of experienced companies to whom we could present our problem and who would give some thought and advice on what method to use in the particular case at hand. Serious geophysical companies will not concentrate readily on small-scale groundwater work. In regional groundwater projects, however, of some larger scope and some prospects to further continuation of the work, serious geophysical contractors are more likely to become interested. This is one of the most serious problems with which we are confronted and I would welcome any suggestions from you how we could go about interesting serious geophysical contractors to give more attention to what is needed in groundwater work both on a regional and a local scale.

Giret, R., Compagnie Générale de Géophysique, France. If I could answer, I would like to say that I am not here to advertise my company, but Compagnie Générale de Géophysique has done more in groundwater geophysics during the last ten years in French-speaking Africa and also in other places, we attempt to understand groundwater exploration by geophysics. I don't think, as you say, that groundwater geophysics does not interest the larger geophysical companies. It is certainly true if you compare groundwater and big volume surveys. Even in our company who does the most in this field, 90% of our work is involved in groundwater. But why? It is only when geologists are aware of what geophysics can give them. I am basically a geologist and I can say that one of the dangers of groundwater exploration with geophysics is that the geologists do not believe that the geophysicists can give them something that is reliable.

Shiftan, Z.L. I have regard for your remarks but we should be aware of the exceptions to the rule.I think geophysical work in Africa is sponsored to a large extent by bilateral government agreements on technical assistance and cannot be considered strictly as an economic operation. It is a research operation from which not necessarily immediate returns are expected. However, I should not like to overstate the case.

Giret, R. We don't work without getting some money. It is true that you cannot make a large investigation in groundwater geophysics without some kind of large organizational planning and in these countries they are generally asking for some aid, bilateral or multilateral or through the FAO. It is true that it is very difficult to raise the money, yet we cannot work without being paid.

The main point that I want to make is that it is a matter of education of the geologist, I can say that because I am one, because there are very few geologists who consider that they need geophysics. In your country, the situation is quite different because you were obliged to use any possible method to organize your water problem. But the situation is not the same elsewhere. We are astonished by the very slow development of geophysics for groundwater in the United States, where, I would guess, there are some water problems too.

Shiftan, Z.L. May I be permitted one last observation: in many papers presented at this conference, attempts were made to demonstrate that certain geophysical methods are able to determine the position of the groundwater level. I was somewhat concerned about this because a hydrogeologist should have a fairly good idea of the position of the groundwater level on the basis of other considerations. I do not believe that the groundwater level can ever be determined to the precision of a meter by present day geophysical methods, this is not what a hydrogeologist would expect from a geophysicist. If it is a nice byproduct of geophysical work, so much the better, but I don't think it is really important to concentrate efforts on this. What we really do need is the extent and subsurface distribution of aquifers.

Integration of geophysical methods for groundwater exploration in the prairie provinces, Canada

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Abstract. In the Canadian prairies, geophysics has had two main uses as an aid in groundwater exploration: delineation of the preglacial drainage pattern and mapping the location and extent of buried permeable deposits of Pleistocene and Recent age. Delineation of the preglacial drainage pattern may involve mapping the elevation of the bedrock surface or simply tracing geophysical anomalies associated with preglacial river courses. Seismic refraction has commonly been used for the former purpose, resistivity and gravity for the latter. Resistivity has also proved suitable for mapping the location and extent of the buried permeable Pleistocene and Recent deposits.

All three methods depend on the existence of appropriate subsurface physical contrasts. Available density, resistivity and seismic velocity data indicate that these contrasts are absent in some areas and field results support the conclusion that none of the methods should find universal application to groundwater exploration in the western Canadian plains. The data concerning physical properties also indicate that it is possible to map surficial and bedrock deposits on the basis of regional variations in these properties. Knowledge of these variations is useful in assessing the suitability of the various methods and in planning an integrated approach to geophysical exploration. Regional variations in drift and bedrock seismic velocities have been mapped and till densities may also have some significant regional variation. In all other cases, the data are far from adequate for mapping purposes.

Considerable effort has been expended on geophysical exploration for groundwater in the Canadian prairies but comparatively little on critical evaluation of the effectiveness of the various methods in a variety of geologic situations. More case-history studies are required to remedy this deficiency and to provide a more reliable picture of the role of each method as a member of an integrated exploration team. Such case-history studies should feature the routine measurement of fundamental physical properties, using core samples and logging techniques. Only in this way can an adequate catalogue of basic physical data be accumulated.

Résumé. Dans les Prairies canadiennes, l'application de la géophysique à la prospection d'eau souterraine s'est manifestée principalement de deux manières: (1) le tracé du réseau de drainage pré-glaciaire, et, (2) le relevé de l'emplacement et de l'étendue de la sédimentation perméable enterrée du Pléistocène et d'âge plus récent. Le tracé du réseau de drainage pré-glaciaire peut comprendre le relevé de la topographie de la roche en place ou simplement le tracé des anomalies géophysiques liées au cours des rivières pré-glaciaires. Il a été essentiellement fait appel à la séismique réfraction dans le premier cas tandis que le second a surtout vu la mise en oeuvre de la résistivité et de la gravimétrie. La résistivité s'est aussi avérée appropriée au relevé de l'emplacement et de l'étendue des sédimentations perméables enterrées du Pléistocène et d'âge plus récent.

Ces trois méthodes sont basées sur l'existence de contrastes physiques de subsurface appropriés. Les renseignements obtenus sur la densité, la résistivité et la vitesse de propagation des ondes montrent l'absence de ces contrastes en certains endroits, et les résultats obtenus sur le terrain confirment la conclusion suivant laquelle aucune des méthodes ne peut s'appliquer d'une manière universelle à la recherche d'eau souterraine dans les plaines de l'ouest du Canada. L'étude des données se rapportant aux propriétés physiques montre aussi qu'il est possible de procéder au relevé des apports de surface et de la roche en place par l'utilisation des variations régionales de ces propriétés. La connaissance de ces variations est utile à la détermination de l'adaptation des différentes méthodes et à l'établissement d'une technique intégrée d'exploration géophysique. Il a été procédé au relevé des vitesses de propagation des ondes des terrains de transport et de la roche saine; il se peut aussi que les densités des dépôts glaciaires soient sujettes à d'importantes variations régionales. Dans tous les autres cas, les données disponibles sont loin de se prêter à la réalisation de cartes.

En comparaison des efforts considérables qui ont été appliqués à la recherche d'eau souterraine par procédés géophysique dans les Prairies canadiennes, il en a été pratiqué bien peu dans le domaine de l'évaluation critique de l'efficacité des différentes méthodes selon les nombreuses conditions géologiques. Il est nécessaire de procéder à l'étude de cas particuliers plus nombreux afin de remédier à cette insuffisance, et de déterminer avec plus de précision le rôle de chaque méthode en tant que partie d'un système d'exploration intégré. Ces études devraient comprendre la mesure routinière des propriétés physiques fondamentales par l'emploi des techniques des différents types de carottage. C'est la seule manière de se faire un dossier sur les données physiques de base.

Geophysical exploration for groundwater in the prairie regions of western Canada requires penetration to depths that rarely, if ever, exceed a thousand feet. There is little incentive for deeper prospecting because present knowledge indicates that prairie groundwaters commonly become increasingly mineralized with depth. There are exceptions to this generalization but, nevertheless on the basis of potability, practical depth limits for groundwater exploration may amount to only a few hundred feet or less in some areas.

Although some extensive major bedrock aquifers contain potable water, surface geophysical methods are not commonly used for their detection since their extent and thickness can usually be better defined from the available geologic evidence. Minor bedrock aquifers, such as thin coal strata or sandstone lenses of limited thickness and areal extent, can rarely be delineated in this way but neither can such aquifers easily be discovered by geophysical methods.

Thus, geophysics on the prairies has found its principal groundwater application in the search for aquifers in the surficial materials overlying bedrock. These include outwash and stream-trench deposits and alluvial sand and gravel deposits of Tertiary and Recent rivers. Under favorable hydrological conditions, the sands and gravels laid down by present-day rivers and streams can be the most important aquifers; of secondary but still major importance are those laid down by preglacial or interglacial rivers and streams. The interglacial deposits, which in some areas are closely associated with preglacial deposits, constitute significant aquifers in both Saskatchewan and Manitoba (J.E. Wyder, pers. comm.) and recent geologic studies suggest that this may be the case in Alberta as well. There are other, less extensive, permeable

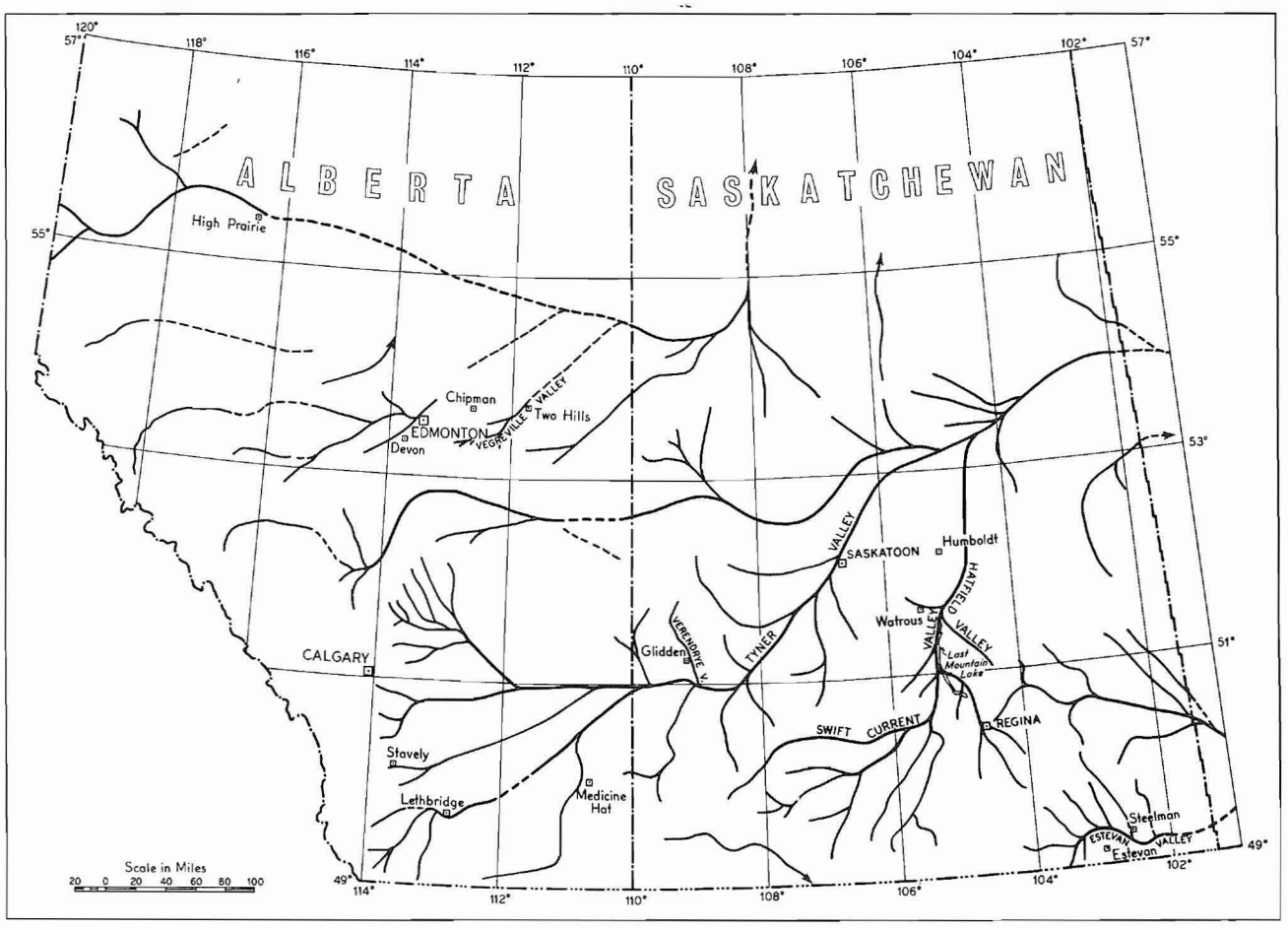


Figure 1. Buried valleys of the western Canadian plains (adapted from Barton, et al. (1964) and from Christiansen (1967)).

deposits in the surficial materials, but generally these are only suitable for domestic, farm, or small municipal supplies. Therefore geophysical exploration of the surficial materials is concerned mainly with the more extensive present-day, interglacial and preglacial deposits and most of the geophysical effort is directed toward the location and tracing of preglacial and interglacial rivers and streams. Figure 1 is a map of the known or suspected courses of these streams based on the maps by Barton, *et al.* (1964) and Christiansen (1967).

Present-day alluvial sands and gravels are readily detected and outlined by the surface resistivity method. The deeper-lying preglacial sands and gravels are generally much more difficult to find by geophysical methods. The approach in this case may be direct or indirect, that is, it may aim at either the direct detection of the aquifer itself, or the detection of a geologic situation favorable for aquifer occurrence. Direct detection is not usually attempted but when it is, the method employed is usually surface resistivity. Although accurate depth and thickness estimates of aquifers are desirable and have been attempted, general practice is to use an almost strictly empirical approach. The usual procedure is the identification of linear high-resistance anomaly trends and their correlation with buried-valley courses. The anomalies in this case are generated by resistivity differences between the permeable deposits and the glacial and bedrock materials in which they are embedded.

The genetic dependence of preglacial sand and gravel occurrence on the preglacial drainage pattern suggests that indirect detection of these deposits is more or less equivalent to delineation of this pattern (Figure 1) or, ideally, to mapping of the bedrock surface. The mapping of this surface, which is now generally concealed beneath a mantle of till and stratified drift, ranging in thickness from a few to about a thousand feet, is theoretically possible by any of the surface geophysical methods commonly used for groundwater exploration on the Canadian Prairies. However in practice the seismic refraction technique has been, almost without exception and despite some serious limitations, the only geophysical method regularly and successfully adopted for bedrock mapping. Surface resistivity and gravity exploration have been used primarily for the delineation of the preglacial drainage pattern by the correlation of anomaly trends with buried-valley courses. The anomalies in this case depend on physical differences between relatively impermeable materials in the preglacial-valley fill and bedrock materials to either side. For resistivity, it may be uncertain whether an observed anomaly is due to resistivity contrasts between galcial sediments and bedrock or to contrasts within the glacial sediments. In other words, it is difficult at first glance to be sure whether one is dealing with a case of detection of a geological situation favorable for aquifer

Table I. Saturated densities of near-surface materials, Alberta and Saskatchewan (gm/cm³).*

Type of Material	Area							
	Glidden, Saskatchewan	Chipman, Alberta	Devon, Alberta	Edmonton, Alberta	High Prairie, Alberta	Lethbridge, Alberta	Stavely, Alberta	Two Hills, Alberta
Surficial								
Clay	2.10 ± 0.04(12)	1.99 ± 0.05(4)	–	–	1.80 ± 0.08(4)	1.94(1)	–	–
Silt	1.98 ± 0.04(34) 2.12 ± 0.05(5)	–	–	–	1.72 ± 0.06(3)	–	–	1.99 ± 0.02(3)
Sand	2.01 ± 0.04(10) 2.22 ± 0.04(5)	1.96 ± 0.06(2)	–	–	–	–	–	–
Till	2.24 ± 0.06(22)	1.96 ± 0.07(26) 2.04 ± 0.05(33)	2.16 ± 0.09(9)	–	2.19 ± 0.07(7)	2.10 ± 0.06(9)	2.22(1)	2.23 ± 0.07(5)
Bedrock								
Sandstone	–	1.98(1)	–	2.04 ± 0.01(4)	–	2.58 ± 0.05(3)	–	2.00 ± 0.06(6)
Siltstone	–	–	–	2.41(1)	–	2.18(1)	–	–
Shale	–	1.75 ± 0.01(2)	–	–	2.16 ± 0.15(5)	2.25 ± 0.06(10)	2.18 ± 0.08(3)	2.06 ± 0.05(2)

*The results presented are the average saturated density, standard deviation and number of observations.

occurrence or with a case of direct aquifer detection. Consideration of the possible effects of geology on the character of the depth-profiling curves may help to resolve the ambiguity.

The effectiveness of geophysical methods in groundwater exploration, as in other applications, increases with relative size of the body to be detected and degree of appropriate physical contrast between that body and the surrounding medium. In the prairie regions of western Canada, direct exploration for buried preglacial sands and gravels runs into difficulty because these deposits are generally relatively thin compared to their depth of burial. On the other hand indirect exploration runs into difficulty because physical contrasts between drift and bedrock are mostly small or nonexistent. This lack of contrast relates in part to the relatively unconsolidated nature of the bedrock and in part to the fact that the drift material is commonly derived from nearby bedrock (Barton, *et al.,* 1964). Thus, densities, resistivities and seismic velocities are in many cases similar and in some indistinguishable.

Gravity exploration

Densities. Table I is a compilation of available density data for samples of near-surface geologic materials in Alberta and Saskatchewan. The Glidden, Saskatchewan results were derived from data presented by Hall and Hajnal (1962); those from Two Hills, Alberta were taken from Lennox and Carlson (1967); the others have been previously unpublished. Hall and Hajnal also obtained an average drift density of 2.00 gm/cm³ by surveying over topographic features.

Hall and Hajnal's measured sample densities are for surficial materials only and were originally presented as logs showing continuous variation of dry and saturated density with depth in two test holes. A summary table showed representative densities for each hole for 50 per cent sample saturation. It is generally accepted, however, that the subsurface is completely saturated below the main water-table (Meinzer, 1923). Although Hall and Hajnal do not indicate water-table depth in either test hole, it is not unreasonable to assume that it lay within several tens of feet or less of the surface. This being so, the bulk of their samples should have been completely saturated *in situ.* Therefore the Glidden, Saskatchewan density statistics in Table I were based on densities read off the saturated-density logs, generally at 5-foot intervals. The intervals were varied slightly where necessary to ensure inclusion in the statistics of all maximum and minimum values on the saturated-density curves. It was felt that this procedure would give a reasonably close approximation to the statistics based on actual sample measurements, since measured densities were for core samples taken at 5-foot intervals.

In three cases, Table I indicates that more than one average density may legitimately be assigned to a given type of geologic material at a given location. In each case two densities are quoted and the density difference between the two is significant at the 0.005 level or better*. In the case of the Glidden silts, relatively low-density silts were found in one of the two test holes, relatively high-density silts in the other; for the Glidden sands, a high-density sand layer was found in between but not contacting two low-density sand layers. At Chipman, Alberta brown till with a density of 1.96 gm/cm³ overlay grey till with a density of 2.04 gm/cm³.

Bedrock densities in Table I are all for Upper Cretaceous sediments. No attempt has been made to assign these values to specific formations, because there is in most cases an element of uncertainty in formation identification. There are no bedrock densities available for the Glidden area, in which the uppermost bedrock is the Bearpaw shale (Hall and Hajnal, 1962). Stearns (1927) has reported a saturated density of 1.98 gm/cm³ for a siltstone sample from the Bearpaw in Montana and one of the Chipman shale samples with a saturated density of 1.74 gm/cm³ is probably also from the Bearpaw.

Average densities for the materials listed in Table I generally vary appreciably with location. There are two apparent exceptions, however: the sandstone samples from Chipman, Edmonton and Two Hills, and the till samples from Glidden, High Prairie, Stavely and Two Hills. Average densities for the three sets of sandstone samples and for the four sets of till samples are 2.01 ± 0.05 gm/cm³ and 2.23 ± 0.07 gm/cm³. The sandstone samples are from neighboring areas and it cannot be inferred that this

*In other words, according to Student's *t*-test (Snedecor, 1956, p. 91) the chance was less than 0.005 that the two density samples being compared were drawn from the same population.

relatively uniform density is characteristic of shallow Upper Cretaceous sandstones throughout more extensive regions of the western Canadian plains. The till samples, on the other hand, are from four widely separated locations and their uniform density may be representative of till density over a considerable area of Saskatchewan and eastern Alberta. With the exception of the Stavely area, the more westerly portions of Alberta seem to be characterized by somewhat lower till densities.

Consideration of Table I suggests that gravity anomalies associated with buried valleys in Saskatchewan and Alberta may be either positive or negative. Where the buried preglacial-valley fill is mainly high-density till and bedrock of the low-density type such as that found at Two Hills or Chipman, a positive anomaly should be observed; where the valley fill is low-density clay, silt, sand or till and the bedrock a high-density type, there should be a negative anomaly. The latter condition is probably common in Saskatchewan (J.E. Wyder, pers. comm.) and holds for some buried valleys in Alberta as well. Of all of the locations listed in Table I, Lethbridge has the most pronounced contrast between bedrock and surficial materials, and this area should be the most favorable for gravity prospecting. Finally, there are a number of combinations of bedrock and surficial materials for which there is little or no density contrast and for which, therefore, no gravity anomalies are to be anticipated.

Results. Application of the gravity method to groundwater exploration in the Canadian prairies has been primarily for the delineation of buried-valley courses (Figure 1). Hall and Hajnal (1962) presented results for the Verendrye and Tyner Valleys of southern Saskatchewan. Robinson (1963) discussed gravity results obtained in the vicinity of the buried Estevan Valley, also in southern Saskatchewan. In both cases, linear Bouguer anomaly trends could be related to the known or suspected courses of the buried valleys. Wyder (1965) subsequently duplicated part of Robinson's survey but disagreed with his findings. Lennox and Carlson (1967) were unable to correlate Bouguer results with a buried-valley system (the Vegreville Valley) north of Two Hills, Alberta. Similarly, it has not been possible to correlate the gravity data for the High Prairie region of Alberta with the available geological information concerning a buried valley passing through the region. However in southern Alberta in the Lethbridge area, there is a possible correlation between gravity results and the course of the buried Lethbridge Valley, although further investigation is required to establish the correlation with certainty.

Hall and Hajnal (1962) observed a number of linear anomaly trends – some negative and some positive – on the Bouguer anomaly map for southern Saskatchewan (Saskatchewan Department of Mineral Resources, 1959). Some appeared to trace out a drainage pattern. Of these, two negative trends were singled out for more intensive study and some detailed gravity traverses were run across them. The resulting residual negative anomalies ranged in magnitude from about 0.7 to about 1.3 mgal. For two of the detailed traverses which crossed the buried Verendrye Valley – a glacial meltwater valley subsequently described by Christiansen (1965) – Hall and Hajnal presented interpreted buried-valley cross sections based on the gravity results. In arriving at these interpretations, they used density contrasts of 0.27 and 0.37 gm/cm^3 between the low-density fill and the higher-density drift or bedrock materials to either side. In both cases the maximum valley relief was estimated to be 400 feet. Hall and Hajnal's cross sections are in reasonably good agreement with those which can be inferred from Christiansen's bedrock topography map. Valley relief for the latter appears to be possibly 50 to 100 feet less but this difference may be due to the lack of geologic control for the deeper parts of the valley or, perhaps, to faulty density assumptions.

Hall and Hajnal also derived an interpreted geologic cross section for a gravity traverse across the preglacial Tyner buried valley, also described by Christiansen (1965). Their most satisfactory interpretation was based on the assumption of a low-density valley fill ranging up to 250 feet in thickness with a density contrast of 0.27 gm/cm^3. In this case, the interpreted cross section is not given and no comparison can be made with Christiansen's map.

All of the gravity anomalies studied in detail by Hall and Hajnal were negative. In explanation, Hall and Hajnal postulated the valley fill to consist principally of low-density silts and the valley flanks of higher-density surficial or bedrock materials. Silts are known to fill the Verendrye Valley at least at one location: a test-hole site near Glidden, Saskatchewan. Silt samples from this test hole (located at the peak of the gravity anomaly) have an average saturated density of 1.98 gm/cm^3; those from a nearby test hole on the flank of the Verendrye Valley an average saturated density of 2.12 gm/cm^3 (Table I). Christiansen's studies indicate the Verendrye Valley to be deeply incised into bedrock cutting through the Bearpaw and Oldman formations into the Grizzly Bear Member of the Belly River formation. Since the density information cited above for the Bearpaw formation indicates its density to be the same or somewhat less than that of the low-density silts, the negative anomaly over the Verendrye Valley possibly depends on relatively high densities in the Oldman and Grizzly Bear strata, for which there are no density data available.

Robinson (1963) continued Hall and Hajnal's studies, investigating the gravity field in the vicinity of the buried Estevan Valley of southeast Saskatchewan (Hall, 1960; Christiansen, 1967). A weak and possibly questionable anomaly appearing on the Bouguer anomaly map for southern Saskatchewan (Saskatchewan Department of Mineral Resources, 1959) coincides in some locations with the suspected course of the buried Estevan. The interpreted anomaly in this case is positive and ranges in size up to about 0.5 mgal. However over most of its course it is appreciably smaller and it disappears completely in some of the more eastern areas.

Comparison of his results for the other detailed gravity profiles with available geologic information led Robinson to postulate that where the positive anomalies existed, they were probably caused by relatively dense materials in the preglacial-valley fill. He concluded that the sand and gravel deposits must commonly be appreciably denser than the nearby bedrock formations. The absence of an anomaly over the easternmost Saskatchewan sections of the buried Estevan was attributed to a termination of the beds of relatively dense sediments.

Wyder (1965) conducted gravity surveys in the areas around Outram and Steelman, Saskatchewan. It was near Steelman that Robinson found his positive anomaly to decrease in magnitude and eventually disappear. Wyder's results also indicated that, at least in some parts of the Steelman area, there was no significant gravity anomaly associated with the buried Estevan Valley. In

other parts of this area, however, where there was an anomaly, Wyder found it to be negative rather than positive and a negative anomaly was found in the Outram area as well. In the Steelman area the anomaly ranged up to about 0.2 mgal in magnitude and subsequent stratigraphic drilling suggested that it was due to thick sand and gravel deposits in the valley fill. In this case it would be necessary to assume that the density of these deposits was appreciably less than that of the nearby bedrock.

Wyder offers no explanation for the differences between his and Robinson's results. However, the interpretation of these anomalies depends heavily on the proper choice of a regional gravity correction. Regional gravity gradients reach 7 mgal/mile or more in this area, whereas the maximum anomaly slopes do not exceed 1 mgal/mile.

Lennox and Carlson (1967), in an intensive geophysical study of an area north of Two Hills, Alberta failed to find any correlation between the Bouguer gravity results and the known course of the buried preglacial Vegreville Valley (Farvolden, 1963). However in this case valley relief was generally less than 200 feet and there were appreciable amounts of silt in the valley fill. Thus, not only did the potential anomaly-producing bodies have a relatively small area of cross section, but also their density contrasts with the neighboring bedrock (Table I) were probably small. Both factors presumably contributed to the absence of any detectable anomalies.

Gravity surveys have also been conducted at High Prairie and Lethbridge in Alberta. At High Prairie, there is generally 300 feet and more of buried-valley relief but, although there is some low-density silt near the surface, the valley fill appears to be mainly till of about the same density as the bedrock shale (Table I). Hence, no gravity anomaly was expected and none was encountered. At Lethbridge, however, there does appear to be an anomaly which correlates with known buried-valley locations (Geiger, 1965) and which is negative, as predicted on the basis of the density data (Table I). According to the available geological information, the valley fill in this case seems to be predominantly till.

Summary and comments. The available information for the Canadian prairies concerning densities of surficial materials and of the uppermost bedrock deposits (Table I) indicates that density contrasts between the two groups of materials may be sufficiently large for measurable gravity anomalies to be associated with buried preglacial valleys; they may also be small enough for anomalies to be either absent or too small to be detected. Detectable anomalies may be either positive or negative.

Detailed gravity surveys to date have revealed a preponderance of negative rather than positive anomalies over known buried valleys. These range in magnitude up to about 1.3 mgal. Some buried valleys or portions of buried valleys have no measurable anomalies associated with them. There is only one reported instance of a positive anomaly and, although two sources support this result, a third reports small negative anomalies over parts of the same valley. No explanation has been offered for this discrepancy but it does suggest the importance of carefully controlled field and data-reduction techniques.

Thus, gravity exploration seems to be a useful but not infallible aid in the location of buried-valley courses on the Canadian prairies and, hence, in the search for groundwater in these areas. Although the results of detailed gravity surveys indicate a preponderance of negative buried-valley anomalies, the available density measurements suggest that positive buried-valley anomalies may not be uncommon, and the geophysicist should be prepared for them. He should also be prepared for the complete absence of anomalies, a phenomenon well substantiated by both available density data and reported survey results. In many cases, he may find it advisable or even essential to supplement his gravity survey with one or more other geophysical techniques to reinforce conclusions based on the gravity results or to obtain information where the gravity method fails to provide it.

Resistivity exploration

Resistivities. Gendzwill (1964) has made the most precise published evaluation of near-surface resistivities in the Canadian prairies. His analysis of the electric logs for oil and potash wells in the Last Mountain-Humboldt area of southern Saskatchewan indicated that most resistivities for the uppermost bedrock stratum in this area – the Marine Shales Formation – are somewhat lower than resistivities for the overlying tills and stratified drift. Although the two sets of resistivities overlapped slightly, about 90 per cent of the shale resistivities ranged between 3 and 6 ohm-m, and about 95 per cent of those for the surficial materials between 9 and 30 ohm-m. The surficial-material resistivity histogram had two distinct maxima: the first at about 15 and the second at about 25 ohm-m. Gendzwill believed these to be related to two distinct classes of drift conditions. Subsequent unpublished electric-log analyses for other areas in Saskatchewan have tended to substantiate Gendzwill's results (J. Dyck, pers. comm.).

Hobson and co-workers (1962) also derived some resistivity values for southern Saskatchewan, in this case for the Estevan area. Analysis of surface resistivity results gave resistivities of 6 to 8 ohm-m, 12 ohm-m, and 15 to 20 ohm-m for the Tertiary Ravenscrag formation, glacial till and glacial outwash material in the buried valley, respectively.

Lennox (1963a) summarized the resistivity data from the analysis of some Wenner apparent-resistivity curves for stations distributed throughout southern Alberta. These analyses suggest a general similarity between resistivity conditions in southern Alberta and those observed by Gendzwill and by Hobson and his co-workers. In both cases bedrock shales appear to have a generally lower resistivity, although the values calculated for Alberta range as high as a few tens of ohm-m. Lennox's southern Alberta results also include values of up to 150 ohm-m for dry surface sands and 500 to 2,000 ohm-m for near-surface gravels.

More recently, Lennox and Carlson (1967) have found that, at least in some parts of Alberta, Wenner apparent-resistivity curves cannot be relied on for the accurate determination of true resistivities. In the Two Hills area, in particular, the Wenner curves commonly indicate bedrock shales to have a lower resistivity than the overlying glacial drift. However, single point electric logs for nearby test holes, show no noticeable resistance change at the drift-shale interface. Burke and Wyder (1967) have remarked the same phenomenon in Saskatchewan and ascribe it to polarization effects in the glacial materials giving rise to a dependence of measured resistivity on current density. Under these conditions, since current densities will decrease at larger electrode spacings the Wenner apparent-resistivity curve for an electrically homogeneous material will resemble the theoretical

curve for a two-layer case, with the upper layer having the higher resistivity. Thus in areas where there is no resistivity contrast at the drift-bedrock interface, the resistivity method may appear to detect the bedrock surface and might be used mistakenly to determine a spurious bedrock depth; on the other hand in areas where there is a resistivity contrast, calculated resistivities and bedrock depths will be somewhat suspect. For these reasons, it is felt that Gendzwill's (1964) electric-log resistivity estimates give the most reliable published values to date for the Canadian prairies.

Although Lennox and Carlson (1967) ran electric logs in the Two Hills area of Alberta, these were all of the single-point resistance type and could not be used to derive true resistivities. Since their surface resistivity measurements also could not be relied on for accurate resistivity estimates, they do not quote any resistivity figures. They do, however, cite a number of resistance changes which, on the basis of their electric-log evidence, seemed to characterize the transition from the surficial materials to the sands and shales of the underlying Upper Cretaceous Belly River Formation. These characteristic changes are listed in Table II.

On the basis of this resistivity information, many of the buried preglacial valleys in the Canadian prairies should contain relatively resistive material and should therefore reveal themselves at the surface as resistivity highs. Some resistive fill will consist of appreciable thicknesses of sand and gravel, others simply of high-resistivity glacial drift. Where the sands and gravels are relatively thin, however, and there is no significant resistivity contrast between the rest of the valley fill and bedrock, no anomaly can be expected. Finally, although resistivity lows should be uncommon, they may be anticipated wherever a drift-filled valley cuts into a sandstone or other relatively high-resistivity bedrock.

Table II. Characteristic resistance changes at the drift-bedrock interface* (Two Hills, Alberta).

Bedrock Material	Surficial Material	Change with Depth at Interface
Sandstone	Clay, silt, or till	Increase
	Sand or gravel	Generally a decrease
Shale	Clay, silt, or till	Little or none
	Sand or gravel	Decrease

*Observed on single-point resistance log.

Results. Resistivity has been widely applied to groundwater exploration in the Canadian prairies. Applications have been principally to the detection and delineation of shallow present-day sands and gravels and to the outlining of buried-valley courses. There has also been one investigation (Wyder, 1963, 1964) of an extensive permeable glacial outwash deposit in the Winkler area of Manitoba. This deposit is about 15 miles long and from 1/8 to 2 miles wide and constitutes a major fresh-water aquifer in the area.

Field equipment has generally been of the direct-current type and the electrodes have been commonly arrayed in the Wenner configuration. However there has been considerable use of the three-electrode and Schlumberger configurations, especially in Saskatchewan.

Application of resistivity prospecting to the detection and delineation of shallow present-day sands and gravels is generally quite routine and, although there are a few brief references in the literature to this application in western Canada (Lennox, 1963b; Wyder, 1963, 1964; Lennox and Carlson, 1967), reports concerning this type of study have generally been restricted to internal circulation within the survey agencies. On the other hand, buried-valley exploration by the resistivity method is not usually so straightforward. A buried valley does not invariably generate a resistivity anomaly and, if it does, the anomaly is not necessarily exactly centered over the valley. Consequently there have been quite a few publications (Anon., 1962; Hobson, *et al.*, 1962; Gendzwill, 1964; Lissey and Wyder, 1966; Lennox and Carlson, 1967) exploring in some detail the problems and results of this application, as well as a series of reports (Wyder, 1964, 1965, 1966) very briefly outlining results.

The buried Estevan Valley of southern Saskatchewan (Hall, 1960; Christiansen, 1967) has been a particularly favored target for resistivity exploration. The Saskatchewan Research Council (Anon., 1962) conducted a number of early surveys comprising nearly 400 depth-sounding resistivity stations and covering most of the buried Estevan in Saskatchewan. At stations located along or near the inferred buried-valley axis, apparent resistivities for large electrode spacings (generally 400 feet or more) ranged up to and over 15 ohm-m; at points remote from the buried valley they were as low as about 10 ohm-m. The author of this report states: "With the exception of an area extending about 10 miles west from Steelman, the courses of the preglacial Estevan and Yellowstone have been well defined for nearly 100 miles across southern Saskatchewan." The course of the buried Estevan, as defined by these early resistivity surveys, however, differs appreciably in the Steelman area from that outlined on the basis of available geologic data by Christiansen (1967). These early results were used by Hall (1960) in arriving at his interpretation of the course of the buried Estevan Valley.

Hobson and others (1962) and Wyder (1964, 1965, 1966) have also carried out resistivity surveys over the inferred locations of the buried Estevan. Hobson and his group concluded that the location of the valley could be easily determined wherever it contained appreciable thicknesses of sand and gravel. At such locations the resistivity results indicated the presence of a highly resistive zone at depth. However where sand and gravel deposits were absent or relatively thin, difficulty was experienced in establishing the valley location. Wyder's resistivity studies were in the Oxbow (1964) and Steelman (1965, 1966) areas. At Oxbow the method appears to have been successful in outlining the buried Estevan and a major tributary; at Steelman it delineated a large resistivity high, attributed to sands and gravels of the buried Estevan system. A previously unreported sand and gravel deposit near Steelman was found in this way and its presence confirmed by subsequent test drilling.

Application of the resistivity method to buried-valley exploration in the Canadian prairies has not been confined to the delineation of the buried Estevan and its tributaries. The buried preglacial Hatfield and Swift Current valleys of southern Saskatchewan have also received considerable attention. Hobson and others (1962) conducted a brief early survey in the vicinity of the Hatfield Valley near Watrous and Gendzwill (1964) subsequently

carried out an extensive survey over a considerable length of this valley as well as over part of the Swift Current Valley. Hobson and his co-workers failed in an attempt to correlate quantitative resistivity interpretations with geology, but geological data were admittedly sparse. Gendzwill found that the resistivity contour maps for large Wenner electrode spacings (600 to 1760 feet) exhibited prominent linear high-resistivity trends which more or less parallel the Hatfield Valley as located on the basis of available geological information. The 1200-foot resistivity contour map presented the most clear-cut picture of the anomaly trend. For this electrode spacing the anomaly has a maximum value of about 17 ohm-m; apparent resistivities away from the anomalous zone were as low as about 6 ohm-m.

Gendzwill believed his anomalies to be caused by the additional thicknesses of relatively high-resistivity drift to be found in the buried-valley locations. This explanation differs from that usually advanced in which the resistivity high over a buried valley is attributed to high-resistivity sand and gravel deposits in the valley fill. Gendzwill's explanation seems quite reasonable on the basis of his electric-log estimates of bedrock and drift resistivities which were cited in the preceding section. However it fails to account for the apparent absence of an anomaly over the Swift Current Valley where it joins the Hatfield unless this area is one of unusually low drift resistivity.

In the southern part of his study area Gendzwill found that the axis of the anomaly trend, wherever it was well defined, was always within a mile or so of the geologically determined Hatfield Valley axis. In the northern part of the area, however, the two axes diverged widely, the extreme separation being as great as 10 miles. There is some possibility that a greater density of geological control in the northern part of the area might identify a tributary valley as the cause of the anomaly trend. Gendzwill points out that five drill holes located along the anomaly in this area encountered exceptionally thick drift, a strong indication that there is an unmapped buried valley underlying the anomaly.

There have been two other applications of the resistivity technique to buried-valley exploration in the Canadian prairies. Lennox and Carlson (1967) concluded from their studies in the Two Hills area of Alberta that high-resistivity anomalies may be associated with buried-valley courses if the valleys contain appreciable amounts of sand and gravel. Lissey and Wyder (1966) worked in the Oak River Basin area near Strathclair, Manitoba which is characterized by highly mineralized groundwaters. In this area the till is less resistive than the underlying siliceous Odanah shale which, in turn, is less resistive than the sands and gravels found in a buried valley running through the study area. Consequently, the observed anomaly was in some instances a high and in others a low, its nature depending on the thickness of sand and gravel in the valley and the electrode spacing.

In almost all of the investigations mentioned, resistivity interpretation has been qualitative, that is, it has involved the location of anomalous trends or areas and not the determination of layer depths and thicknesses. There is fairly general agreement among observers (Lennox, 1963b; Gendzwill, 1964; Lissey and Wyder, 1966; Lennox and Carlson, 1967) that determination of depths and thicknesses in the typical Canadian prairie geological environment is difficult or impossible because of the generally complex nature of the drift and, in some cases, of the bedrock sections. This applies in particular to those methods which assume horizontal homogeneous isotropic layering. It also applies to Gendzwill's empirical method of bedrock depth determination for the Last Mountain-Humboldt area which is based on statistical correlation of measured surface resistivities and known bedrock depths. Gendzwill estimates that the possible error in depth determination by this method is about 100 feet for depths ranging up to about 900 feet. The accuracy may improve where bedrock is relatively shallow.

Summary and comments. Resistivity measurements for surficial materials and for the uppermost bedrock deposits in the Canadian prairies indicate that glacial drift and bedrock shales and sandstones are less resistive than the alluvial sands and gravels deposited in buried valleys by preglacial and interglacial streams. Although there is considerable evidence to suggest that bedrock shales are commonly less resistive than the glacial materials, there is at least one case of a shale that is more resistive than the nearby till, and in some areas there may be no resistivity contrast between the two. However polarization effects may cause surface resistivity measurements in the latter areas to resemble those for a layered situation. There are few resistivity data for bedrock sandstones but those available show one of the sandstone members of the Belly River formation to be more resistive than the overlying drift.

The resistivity data suggest that buried preglacial valleys will commonly generate resistivity highs, although in some instances there may be either a resistivity low or the absence of any anomaly. Field results have confirmed this hypothesis. Almost all the observed anomalies have been highs, although there has been one reported low and one case in which no anomaly was observed for a distance of approximately 10 miles along the presumed course of a known buried valley.

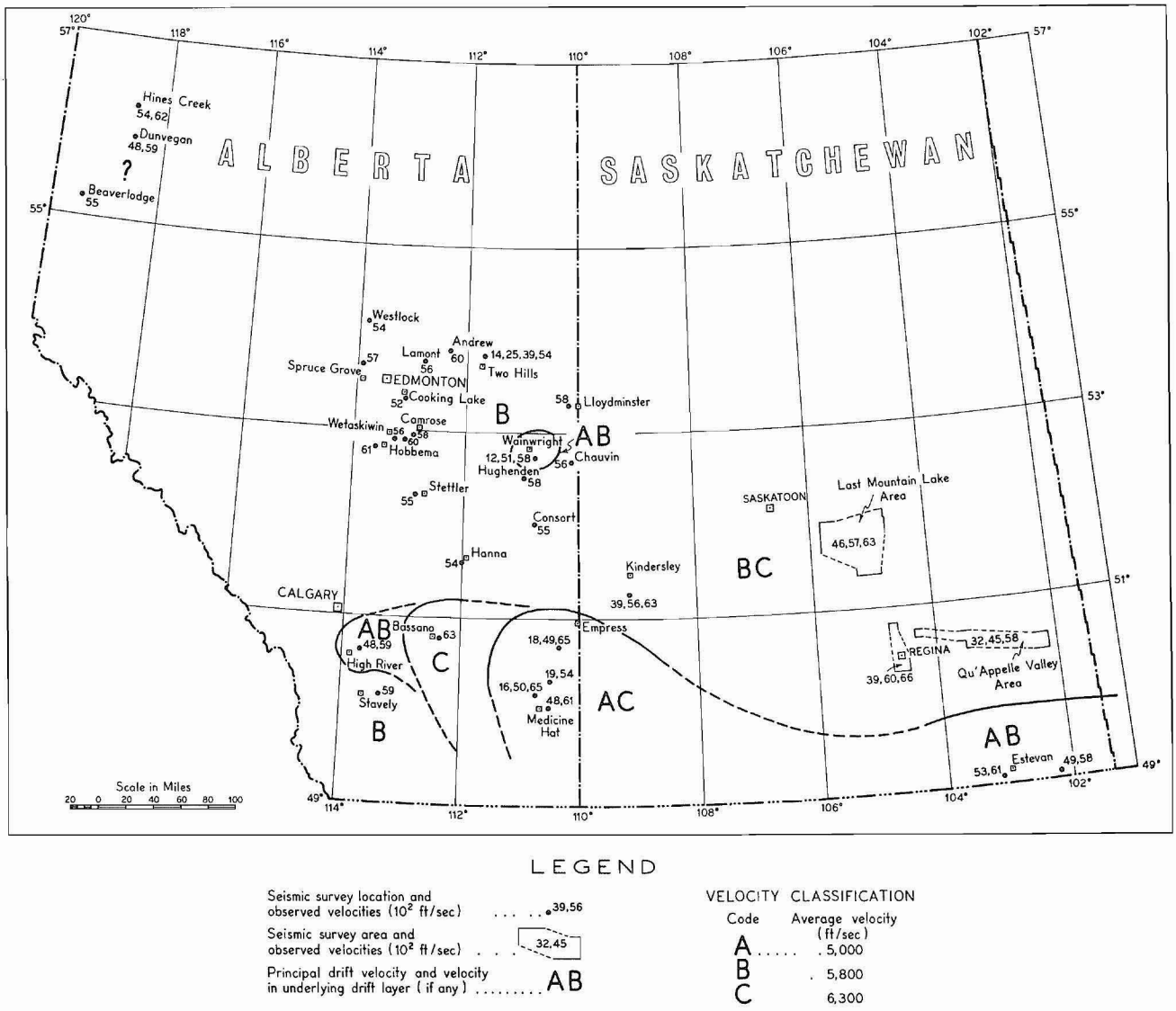
In at least one case resistivity results have led to a modification (Hall, 1960) of a buried-valley course originally determined on the basis of available geological data. It appears to be generally more reliable than the gravity method as a buried-valley indicator, although there is still some possibility that some buried valleys will fail to generate an anomaly. Where resistivity anomalies are small or absent, it may be necessary to call upon other geophysical methods.

Table III. Horizontal seismic velocities, Greater Winnipeg Floodway area (Hobson, *et al.*, 1964).

Type of Material	Velocity Range (ft/sec)
Surface layer, loam, silt, etc.	<1,500
Sandy clay	2,000 - 3,800
Clay	2,000 - 6,000
Unconsolidated till	6,000 - 7,400
Semiconsolidated till	7,050 - 9,050
Consolidated till	9,800 - 13,000
Bedrock (limestone or dolomite)	>12,200

Seismic exploration

Velocities. Figures 2 and 3 are maps summarizing available horizontal seismic velocity information for surficial materials and bedrock, respectively, in Saskatchewan and Alberta. Each velocity shown is the average of a relatively narrow range of velocities observed in the area outlined, or in the vicinity of the



LEGEND

Seismic survey location and observed velocities (10^2 ft/sec) 39,56

Seismic survey area and observed velocities (10^2 ft/sec) . . . 32,45

Principal drift velocity and velocity in underlying drift layer (if any) . . . . . AB

VELOCITY CLASSIFICATION

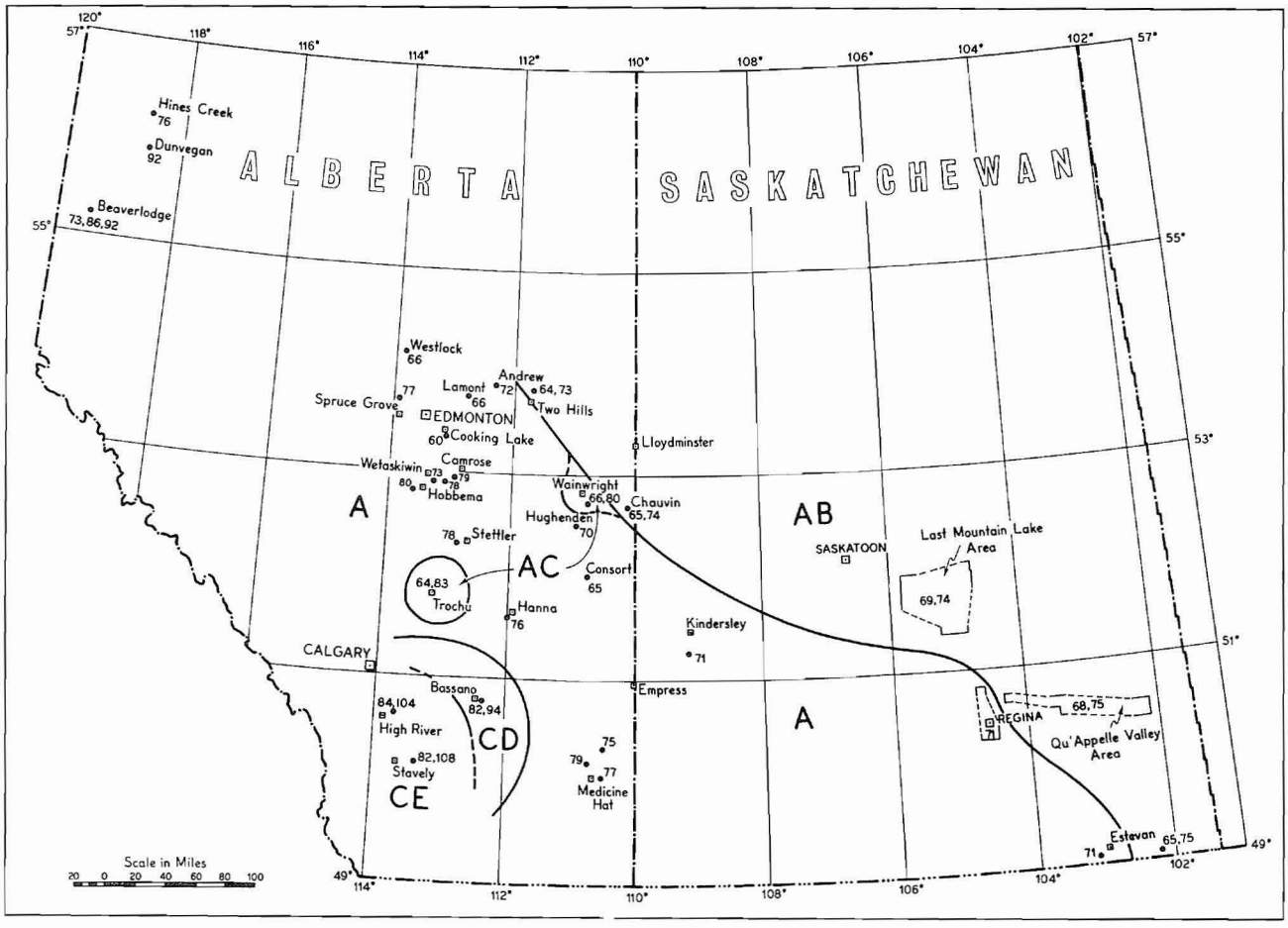
Code	Average velocity (ft/sec)
A	5,000
B	5,800
C	6,300

Figure 2. Surficial-material velocity map.

location indicated. Calculated standard deviations average about 300 ft/sec. Velocities have also been obtained for the Winnipeg area of Manitoba using the hammer seismic method (Hobson, *et al.*, 1964). However these velocities (Table III) do not seem to relate readily to those shown in Figures 2 and 3. The differences are indicative of the more detailed velocity information commonly obtained with the hammer seismic method, as well as of a difference in geological settings. Bedrock in the Winnipeg area is the high-velocity Ordovician Red River formation and the local tills consist of carbonates and Precambrian materials. Bedrock in the other surveyed areas of Alberta and Saskatchewan is low-velocity Cretaceous and Tertiary shale or sandstone and the local tills consist primarily of Cretaceous shale materials.

The Saskatchewan glacial drift and bedrock velocities shown in Figures 2 and 3 were calculated from the velocity information given by Hall (1962) on his seismic cross sections for southern Saskatchewan. Hall's results were based on the analysis of first arrivals on oil company seismic records. The Alberta velocities summarize most of the near-surface results obtained to date in the province by the Research Council of Alberta. Velocities for the seismic aerated surface layer and all hammer seismic velocities have been omitted. In the case of the seismic aerated layer, which in Alberta commonly ranges in thickness from about 5 to about 15 feet and in velocity from about 500 to about 1000 ft/sec, the reason for omission is primarily a lack of precisely determined values; the Alberta hammer seismic velocities generally do not correlate readily with velocities obtained by conventional prospecting or with known lithologies. Alberta hammer seismic velocities commonly suggest the existence, within the first few tens of feet of the surface, of a number of thin layers with discrete seismic properties or of a zone in which seismic properties change rapidly but continuously with depth. The conventional refraction methods do not yield this detailed type of near-surface velocity information.

Seismic velocity histograms for drift and other surficial materials are presented in Figure 4. The first histogram is for



LEGEND

Seismic survey location and observed velocities (10^2 ft/sec) . 65,75

Seismic survey area and observed velocities (10^2 ft/sec) . 69,74

Principal bedrock velocity and velocity in underlying bedrock layer (if observed) . . AB

* Z = depth to top of layer

VELOCITY CLASSIFICATION

Code	Average velocity (ft/sec)
A	4,800 $Z^{0.086}$ (Alta.) * 5,100 $Z^{0.060}$ (Sask.) *
B	7,400
C	8,300
D	9,400
E	10,700

Figure 3. Bedrock velocity map.

Table IV. Horizontal seismic velocities, Alberta and Saskatchewan (ft/sec)*.

Type of material	Area		
	Alberta	Saskatchewan	Combined
Unsaturated sand(?)	1,200 − 2,100(32)	–	1,200 − 2,100(32)
Alluvium, silt, gravel	3,600 − 4,300(4)	3,200 − 4,900(15)	3,200 − 4,900(19)
Drift	5,000 ± 400(31)	5,200 ± 200(6)	5,000 ± 400(37)
Drift	5,700 ± 400(251)	5,800 ± 300(113)	5,800 ± 400(364)
Drift	6,400 ± 300(27)	6,300 ± 200(62)	6,300 ± 200(89)
Bedrock	7,100 ± 700(235)**	6,900 ± 300(109)***	–
Bedrock	7,300 ± 300(13)	7,500 ± 400(7)	7,400 ± 400(20)
Bedrock	8,300 ± 400(94)	–	8,300 ± 400(94)
Bedrock	9,400 ± 400(4)	–	9,400 ± 400(4)
Bedrock	10,700 ± 700(77)	–	10,700 ± 700(77)

* The results are presented as ranges, or averages and standard deviations. The figures in brackets are the numbers of observations. ** Can be expressed as $4800z^{0.086}$ where z is depth. *** Can be expressed as $5100z^{0.060}$ where z is depth.

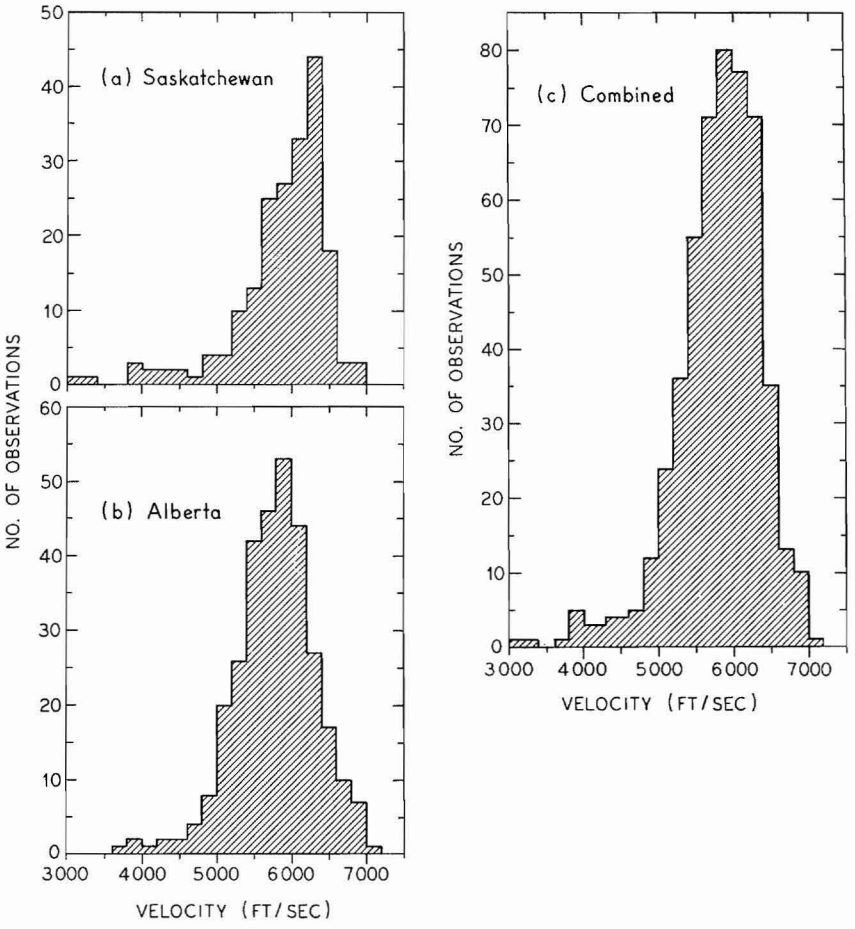


Figure 4. Surficial-material velocity histograms.

Saskatchewan, the second for Alberta, and the third combines data for both provinces. The few velocities less than 3000 ft/sec have been omitted. All three histograms display a single strong peak, the peak velocity being slightly higher for Saskatchewan than for Alberta. A more complex velocity system is suggested by the surficial-material velocity map (Figure 2). Two or more discrete surficial-material velocities have been observed in some survey areas, particularly in southeastern Alberta and throughout Saskatchewan. Furthermore, there appear to be several relatively well-defined areas, each characterized by a distinctive set of drift velocity conditions. On the basis of this evidence for layering within the surficial materials and for characteristic areal velocities, it has been possible to derive three statistically distinct representative drift velocities for the survey areas in Saskatchewan and Alberta. These velocities, together with velocities believed to be typical for Alberta and Saskatchewan of unsaturated sands or other permeable materials and of near-surface alluvium and colluvium, are listed in Table IV. The drift velocity regions shown in Figure 2 were outlined on the basis of the areal distribution of the three representative drift velocities.

There are two major and three minor drift velocity regions. The principal region covers most of the areas surveyed in southern Alberta and Saskatchewan. The uppermost drift layer in this region, with an average velocity of 5800 ft/sec, is underlain in Saskatchewan by a second drift layer with an average velocity of 6300 ft/sec. The second major drift velocity region is found in the extreme southeastern parts of Saskatchewan and Alberta and is assumed to extend across southern Saskatchewan between these two areas. The uppermost drift layer in this region has a velocity of 5000 ft/sec. It is underlain in Alberta by a drift layer with an average velocity of 6300 ft/sec and in Saskatchewan by a layer with a significantly different average velocity of 5800 ft/sec. Each of the three minor drift velocity regions essentially covers only the area of a single seismic survey, although the boundaries have been arbitrarily extended outward to some degree in all three cases.

The characteristic drift velocities and drift velocity regions seen on Figure 2 presumably relate in some way to the glacial history of western Canada and, given sufficient data, might be correlated with the observed sequence of glacial deposits and with the areal distribution of these deposits. The surficial deposits over more than 50 per cent of southern Alberta and Saskatchewan consist of till overlain by lacustrine sands, silts and clays. Within this same region there are large areas of 'stratified till', that is, of till interbedded with lacustrine and ice-rafted materials. Thus there is ample evidence for geological layering in the surficial materials. Whether this layering leads to the observed seismic layering is yet to be demonstrated.

Bedrock velocity histograms are presented in Figure 5. In the case of the surficial velocities there is a histogram for Saskatchewan, one for Alberta and a third combining results for the two provinces. The Saskatchewan histogram shows a strong single

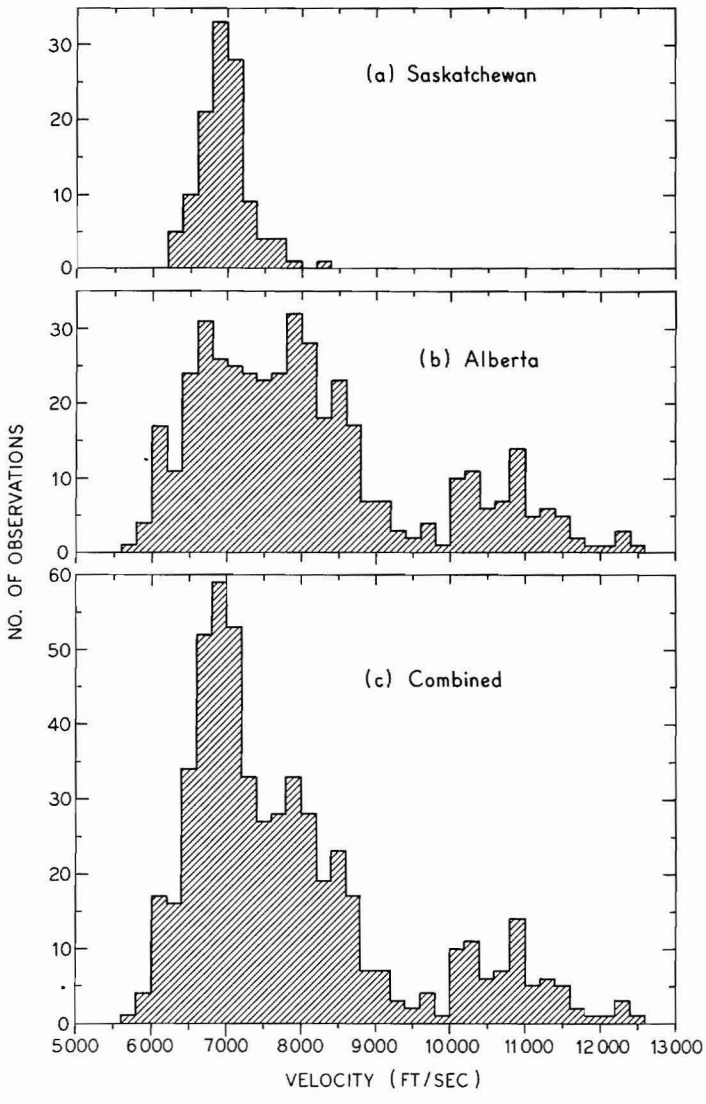


Figure 5. Bedrock velocity histograms.

peak, whereas that for Alberta suggests a complex distribution of velocities. The bedrock velocity map (Figure 3) also suggests a complex distribution, not only for parts of Alberta, but also for some areas in Saskatchewan. As in the case of the drift velocities, the complexity is in part a consequence of layering and in part a consequence of the existence of characteristic areal velocities. Representative and statistically distinct bedrock velocities for the two provinces are listed in Table IV. They form the basis of the subdivision into velocity regions shown in Figure 3.

There are basically two major velocity regions shown on the bedrock velocity map. The first covers most of the areas surveyed in Alberta and Saskatchewan and is characterized by an uppermost bedrock layer whose seismic velocity can be expressed as a function of a fractional power of its depth of burial. This major region can be subdivided into two minor regions. The first covers most of the Alberta survey area and a large southwest portion of Saskatchewan; no deeper-lying bedrock layers have been observed in this subdivision except in the vicinities of Trochu and Wainwright where there is a deeper layer with an average velocity of 8300 ft/sec. The second subdivision comprises a small northeastern section of Alberta and the northeastern portion of the surveyed areas in Saskatchewan; throughout this minor region there is observed an underlying layer with an average velocity of 7400 ft/sec. The second major region is found in the southwestern area of Alberta where the uppermost bedrock layer has an average velocity of 8300 ft/sec. Underlying bedrock layers are observed with an average velocity of 9400 ft/sec in the Bassano area, and 10,700 ft/sec farther to the west. The seismic data in the Stavely area of this region are undergoing re-evaluation and there may be some modification of these velocity findings as a result.

Those bedrock velocities which can be expressed as fractional powers of the depth are of particular interest. Theoretical studies (Iida, 1939; Gassmann, 1951) have shown that seismic velocity in a simply packed granular material should vary as the sixth power of the depth of burial. Field and laboratory studies of sedimentary rocks (Faust, 1951, 1953; Berry, 1959; Wyrobek, 1959; Acheson, 1959, 1963) generally fail to yield the theoretical result predicted by Iida and Gassmann but lead instead to a more generalized velocity-depth relation of the form:

$$V = cz^{1/n} \qquad 1$$

in which V is velocity, z depth, and c and n are constants. The empirical values of n vary from about 4 to about 33.

Derivation of these velocity-depth relations has been commonly based on the analysis of velocity data from well surveys, although Berry's results were for laboratory measurements of sandstone cores. The Alberta and Saskatchewan velocity-depth relations given as footnotes to Table IV are based instead on statistical correlation of the logarithms of the two variables, observed velocity and computed depth to the top of the bedrock layer. In a few locations where true depths were known accurately these were substituted for computed depths. The attempt to correlate was prompted by the apparent tendencies for observed velocities to increase with depth to the refracting layer, as revealed on scatter diagrams. These diagrams are shown in Figure 6 for the principal uppermost bedrock velocities in Saskatchewan and Alberta. The two corresponding computed velocity-depth relations are also shown on the same figure.

Application of Student's t-test (Snedecor, 1956, p. 173) has demonstrated that both of the velocity-depth relations given in Table IV and shown in Figure 6 are statistically significant, that is, that there is a high probability* of a real velocity-depth correlation and that this correlation may be expressed by a relation of the type given by Equation 1. Further application of the t-test (Snedecor, 1956, p. 178) has indicated that the difference between the two computed relations probably reflects some real geological difference (this is also true of the same data expressed in terms of averages and standard deviations). Hence, there has been no attempt to combine both sets of data to obtain a single comprehensive velocity-depth relation for the Canadian plains area.

Similar velocity-depth relations may eventually be developed for the other bedrock layers listed in Table IV, as well as for some

*In each case the correlation was significant at the 0.001 level, i.e., the chance was less than 0.001 that the sample was taken from a population for which there was no velocity-depth correlation.

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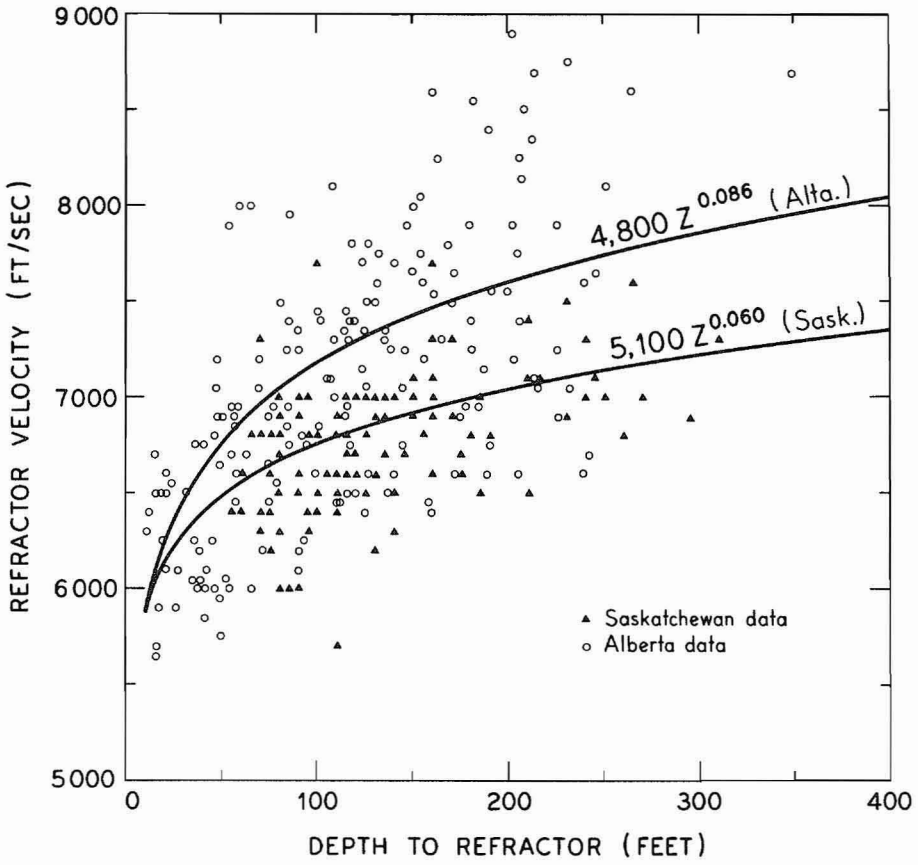


Figure 6. Velocity variation with depth for the uppermost bedrock layers.

of the drift layers. However proof of statistical significance generally requires a relatively large number of data, particularly in a case like this for which velocity is influenced by other factors such as geological age and lithology (Faust, 1953). The relative scarcity of data for the other layers probably explains why it has not been possible to determine significant velocity-depth relations for them.

Some general inferences concerning the size of the drift-bedrock velocity contrast may be drawn from a comparison of the two velocity maps (Figures 2 and 3). Contrasts should be generally good (averaging 2000 ft/sec or more) in the southwest part of Alberta and, purely on this basis, this area should be favorable for bedrock mapping by the seismic refraction method. Contrasts should be generally adequate (averaging 800 ft/sec or more) in the strip extending across the area from the southeast portion of Alberta to the southeast portion of Saskatchewan. Bedrock mapping should be feasible throughout this area, although difficulties could be experienced locally because of smaller than average velocity contrasts. Such difficulties have been noted in the Estevan area (Hall, 1962). For the remaining surveyed areas in the two provinces, difficulties may be expected where bedrock is shallow, because of velocity variation with depth of burial for the uppermost bedrock layer. In Saskatchewan especially, where the drift commonly comprises two layers, adequate velocity contrasts (at least 400 ft/sec or more) may require bedrock depths exceeding 60 feet. In Alberta, with a single-layer drift, the corresponding figure is about 20 feet. These depth estimates were based on a comparison of drift velocities with bedrock velocities as predicted by Figure 6 for various thicknesses of drift cover.

The bedrock velocity data given in Table IV and the corresponding velocity regions outlined in Figure 3 are for layers having an adequate velocity contrast with the overlying drift or bedrock material. Layers with seismic velocities less than, equal to, or only slightly greater than those for the overlying materials will not be detected by the refraction method. Such layers will be suspected, however, wherever computed seismic depths are consistently and appreciably greater than estimated depths based on available geologic control, or wherever bedrock cannot be detected even though it should be found within the depth of penetration of the refraction method. There is some reason to believe that the uppermost bedrock layers in the Wainwright and Lloydminster areas of Alberta have velocities at best only slightly greater than those for the overlying drift since bedrock cannot be detected at Lloydminster and its depth is grossly overestimated at Wainwright.

The discussion so far has concerned horizontal velocities exclusively. If there is any question of seismic anisotropy, it will be important to have some knowledge of vertical velocities as well. Vertical velocities can be determined most accurately by means of well surveys, which provide information not only about anisotropy but also about inadequate velocity contrast, velocity reversals and 'blind zones' (Soske, 1959). Uphole velocity surveys in the Estevan area of southern Saskatchewan (Hobson, *et al.*, 1962) disclosed the presence within the drift of thin high-speed layers in which vertical velocities approaching 10,000 ft/sec were

measured. These layers were too thin to be detected by the refraction method, but they might certainly affect the quality of the refraction results. Vertical seismic velocities have also been surveyed near Two Hills, Alberta (Lennox and Carlson, 1967). They provided evidence indicative of anisotropy in the aerated layer at the surface, but all other layers could be interpreted as isotropic. They also tended to confirm a supposition based on the refraction results for the same area that there was a significant velocity variation with elevation for the bedrock surface.

Results. There has been little published concerning the application of seismic techniques to groundwater or other types of shallow exploration in the Canadian prairies. Most publications on this topic have to do with the refraction technique. However Hobson and others (1962) used both the refraction and the high-resolution reflection method in seismic traverses over the buried Estevan Valley near Estevan, Saskatchewan. For bedrock depths ranging up to 500 feet for the deepest parts of the valley they found that: "Bedrock (could) be identified with some degree of accuracy from several reflection profiles but not as definitely from the refraction profiles." They concluded that the seismic reflection technique was the more suitable of the two for bedrock depth determination in the Estevan area, but recommended the use of the refraction technique and of velocity logging to provide velocity information. Geological control in the Estevan area was sparse at the time of this survey and these conclusions may be subject to revision.

Of publications dealing with the refraction technique, those by Hall (1962) and Lennox and Carlson (1967) do so in some detail. Hobson and co-workers (1964) briefly describe a hammer seismic survey of the Red River Floodway area of greater Winnipeg. Because of the generally excellent velocity contrasts existing in this area (Table III), interpretation of the seismic results was presumably fairly straightforward and there was good agreement between bedrock depths determined by test drilling and by the seismic method. For drift thicknesses ranging from about 45 to about 90 feet, there was an average seismic error of about 7 per cent.

Hall (1962) used the 'first breaks' on oil company seismic records to determine bedrock depths in Saskatchewan by the refraction method. Since these records were originally intended for the determination of deep structure by the reflection method, they were not ideally suited for Hall's purposes, their most serious limitation being lack of penetration. For conventional reflection spread lengths of 1/4 mile or less, maximum penetration by the refraction technique was commonly less than 100 to 200 feet. Despite this limitation and that imposed by the relatively low velocity contrast between drift and bedrock (Table IV), Hall found that most of his seismic cross sections gave useful information. Deposits of sand, gravel, silt and alluvium – generally characterized by low seismic velocities (Table IV) – were readily detected if located at or near the surface and possessing lateral dimensions comparable to seismic spread length. Hall was also able to identify interfaces within the drift and between drift and bedrock. However he noted that usually it was not possible to continue these interfaces right across a given cross section on the basis of the seismic evidence. He offered three possible explanations for the observed discontinuities: existence of a real geological discontinuity, disappearance of velocity contrast due to velocity variation, and depth of the refracting layer exceeding the maximum depth of penetration.

A number of Hall's cross sections were for traverses across known buried valleys, including the Estevan, Swift Current, Verendrye and Hatfield valleys. Seismic bedrock depressions were found or suspected in the vicinity of the buried Estevan near Outram and Steelman, but their locations did not coincide with the presently accepted buried-valley course nor with the locations of associated resistivity and gravity anomalies. Nevertheless the seismic bedrock interpretation in the Outram area was felt to be in general agreement with known geology. For the Swift Current Valley south of Regina, the seismic bedrock surface was judged reliable enough to be incorporated in a report of the geology and groundwater resources of the Regina area (Christiansen, 1961). A pronounced seismic bedrock depression was observed within a mile or so of the geologically inferred axis of the Swift Current Valley in an area south of Humboldt where the Swift Current Valley joins the Hatfield Valley. There was no such clear-cut evidence for the existence of the Hatfield Valley, either on adjoining cross sections or on a line of cross sections a few miles to the north, although both sets of traverses must have crossed the Hatfield at some point. Similarly, there was no seismic indication of a bedrock depression for two traverses across the buried Verendrye meltwater valley, although low-velocity surficial material found at the valley location for one of these traverses was perhaps identical with the low-density silts observed by Hall and Hajnal (1962) at the same location. However the indicated seismic thickness was a great deal less than the silt thickness found during drilling. Finally, for some traverses across and parallel to the present-day Qu'Appelle River, it was felt that there was reasonable agreement between the seismic bedrock depth predictions and available geologic control.

Although Hall concludes that the refraction method can be applied very effectively to certain classes of problems arising in groundwater exploration in Saskatchewan, the results suggest that its performance has been rather erratic. Its inability to detect bedrock or to locate known buried valleys consistently is undoubtedly due to the existence of low velocity contrasts in at least some of the areas surveyed. It was suggested on the basis of the velocity observations (Table IV) that this should prove to be less of a problem for the two surveys near Estevan than for the other four farther north, but the survey results provide no solid support for this hypothesis. One of the Estevan surveys does have a high success ratio measured in terms of the ratio of the number of successful bedrock depth determinations to the number of seismic spreads employed but the success ratio for the other is lower than average.

Lennox and Carlson (1967) carried out an extensive seismic survey in an area of about 40 square miles to the north of Two Hills, Alberta. In addition to conventional spreads of about 1/3 mile in length, the survey featured many short spreads with closely spaced detectors designed to delineate structure very near the surface. Available geologic data were used abundantly and these were supplemented, where necessary, by stratigraphic test drilling. Vertical seismic velocities measured in five of the test holes proved useful for the final interpretation of the seismic results. In addition to a principal drift layer with an average velocity of 5400 ft/sec, four low-velocity surficial-material layers were also detected, with average velocities ranging from 800

ft/sec for the aerated surface layer to 3900 ft/sec for an unidentified material. The uppermost bedrock layer had an average velocity of 6400 ft/sec.

Despite this relatively good contrast between drift and bedrock velocities, some difficulty was experienced in accurately determining the depth to bedrock. This was because of the blind-zone behavior of the principal drift layer and, secondly, the effects of velocity variation with elevation for the uppermost bedrock layer. Soske (1959) used the term 'blind zone' to describe any layer of thickness large enough to be appreciable but small enough to prevent the appearance of its characteristic velocity on the time-distance graph. In the Two Hills area it was found that the blind-zone problem could be dealt with fairly satisfactorily by assuming that where the 5400 ft/sec layer was not observed on the time-distance graphs, it had the maximum possible thickness consistent with its nonappearance. This approach led to acceptable bedrock depths in about two thirds of the cases treated.

With the blind-zone corrections incorporated into the results, computed seismic bedrock depths throughout most of the Two Hills survey area compared favorably with nearby geological control. For some spreads where bedrock was shallow, percentage error was large, but relatively small percentage errors in areas of deep bedrock meant that buried valley courses could still be determined with reasonable accuracy. There were a few isolated areas for which the computed seismic depth was much greater than that indicated by the available geological control. As suggested by the seismic results, bedrock topography for these areas did not conform to the preglacial drainage surface but consisted instead, in each case, of a seemingly closed trough or hollow in a bedrock high. The restriction of these anomalous areas to regions of otherwise elevated bedrock suggested the possible existence of a velocity variation with elevation for the uppermost bedrock layer. Such a variation could be a consequence of variation with geological age which Faust (1953) has shown to be valid for deeper-lying sediments or it could be simply a consequence of depth of burial or a combination of these two factors. In regions of elevated bedrock, velocity variation with elevation could lead to inadequate velocity contrasts at the drift-bedrock interface, or even to velocity reversals. Under these conditions, the true bedrock surface would not be detected. The apparent bedrock surface would actually represent some deeper-lying relatively high-velocity bedrock layer.

There was some supporting evidence for the existence of unsuitable velocity contrasts at the drift-bedrock interface in the anomalous areas. One of the refraction records from one of these areas exhibited characteristics demonstrated by Press and Dobrin (1956) to be typical of records taken over a section consisting of a high-velocity surface layer underlain by lower-velocity material. Borehole velocity measurements in one of the anomalous areas indicated that the uppermost bedrock layer had a velocity about equal to or slightly lower than that for the overlying drift. There was, unfortunately, no practical method for correcting the erroneous bedrock depths computed for regions of elevated bedrock, and doubtful values in these areas could only be rejected.

Lennox and Carlson concluded that the seismic refraction method could be successfully applied in the Two Hills area and probably in other areas of Alberta for the approximate determination of bedrock depth and for tracing the courses of buried preglacial valleys. Successful results depend on the availability of supplementary geological and seismic information obtained from lithological logging and velocity surveying of suitably located boreholes. These supplementary data are required for the resolution of such problems as blind zones, velocity reversals, inadequate contrasts and anisotropy. Such problems probably exist to a greater or lesser degree throughout Alberta and will become apparent when the results of detailed refraction surveys are compared with good geological and geophysical control. Some difficult areas are already known. They include the Lloydminster and Wainwright areas where there is a suspected lack of velocity contrast at the drift-bedrock interface and the Stavely area of southwestern Alberta where the uppermost bedrock layer is believed to constitute a blind zone. These difficult areas may provide interpretive methods which will be useful in other areas.

Summary and comments. The available near-surface seismic velocity data for Alberta and Saskatchewan indicate that the velocity contrast between drift and bedrock will be good in the southwestern part of Alberta, fair in a strip extending from the southeastern portion of one province to the southeastern portion of the other, and range from fair to poor in other areas of the two provinces in which shallow refraction surveys have been conducted. The chances of determining bedrock depth and mapping bedrock topography by the seismic refraction method will vary from area to area accordingly. Success or failure in the fair-to-poor area will depend to some degree on the amount of drift cover over bedrock since there is a statistically significant velocity-depth correlation for the uppermost bedrock layer throughout this area: where bedrock is shallow the drift-bedrock velocity contrast is probably small; where it is buried at depths exceeding about 60 feet in Saskatchewan or about 20 feet in Alberta the contrast is probably adequate.

The good velocity-contrast area in southwestern Alberta may extend to the northwest in a strip paralleling the foothills and mountains belts. Good velocity contrasts in southwestern Alberta depend primarily on the existence of relatively high bedrock velocities (Figure 3) and these velocities tend to increase as the foothills and mountains are approached (Kaarsberg, 1959; Acheson, 1963).

Refraction surveys in Alberta and Saskatchewan have provided general confirmation of the predictions based on the velocity results. Attempts to map bedrock in Saskatchewan by this method have commonly resulted in discontinuous drift-bedrock interfaces and failure to discover bedrock depressions from the results of seismic traverses crossing known buried valleys. There were successes as well as failures, however, so that the performance to date in Saskatchewan is best described as erratic. A detailed investigation in Alberta, for which large amounts of geological and geophysical control data were available, demonstrated that the method could be used successfully, given adequate control, for the approximate delineation of preglacial topography and for the tracing of buried-valley courses. It was demonstrated that the velocity along the top of the uppermost bedrock layer in the survey area tended to vary with elevation, so that velocity reversals or inadequate contrasts were a problem in parts of the area. The principal drift layer behaved as a blind zone and caused additional difficulties. Adequate geological and geophysical control is essential for coping with

problems such as these, and it is to be assumed that it will be required to some extent in most, if not all, parts of the western Canadian plains.

Results from both provinces indicate that there are at least a few areas in which velocity reversals or inadequate velocity contrasts will prevent effective use of the seismic refraction method for bedrock depth determination. In others, blind zones or anisotropy may lead to difficulties, and these could be sufficiently great to make the method useless in these areas. If these conditions apply, other methods will have to be adopted, in some cases for the reinforcement of conclusions based on seismic work, in others as the only means of obtaining useful information.

Integration

General. Dobrin (1952) has discussed the possible advantages to be gained by utilizing a combination of geophysical methods rather than limiting geophysical exploration to a single technique. There are basically two: reduction in exploration costs and increase in quantity or quality of data. Exploration costs can be reduced by reserving effective but expensive methods for detailed surveys in promising areas, and adopting inexpensive methods for preliminary reconnaissance surveys. Increase in data quantity results if a supplementary geophysical method either confirms the prior discovery by another method of an anomalous subsurface body or structure, or leads to a more reliable picture of the geometry and dimensions of that body or structure. Data quality is increased if a supplementary method points out possible sources of error in an earlier method and provides information that may be useful in correcting these errors.

There are many reported examples of the integration of geophysical methods in the petroleum and mineral exploration fields, and numerous publications concerning the application of a variety of these methods to shallow exploration of the sort common in groundwater prospecting. However there is a comparative shortage of reports dealing specifically with integration of methods in groundwater or related fields. Among the detailed reports of this type that might be cited are those by Ahmad and Thirlaway (1958), Kailasam (1958), and McGinnis and Kempton (1961). Briefer reference to the combined use of two or more geophysical methods for groundwater purposes can be found in publications by Woollard and Hanson (1954) and Kelly (1962).

Ahmad and Thirlaway (1958) used gravity, seismic refraction, seismic reflection and a limited amount of exploratory drilling to determine the groundwater potential of thick alluvial deposits in a mountain valley in western Pakistan. The gravity and seismic reflection methods were used to provide independent estimates of thickness for the valley fill; the velocities from seismic refraction were used in the identification of fill materials. The survey indicated that there were no productive aquifers within the valley. Comparison of the gravity and other data for this valley with similar data for a neighboring valley containing highly productive alluvial aquifers suggested that the first valley was filled with fine-grained lake deposits laid down when its usual outlet was blocked, whereas the second contained thick series of porous gravel and boulder beds deposited by a normal, unobstructed mountain river system.

Kailasam (1958) carried out integrated seismic refraction and resistivity investigations in the Narmada basin in India. He found that it was possible, under favorable conditions, to obtain information regarding depth to bedrock and nature of bedrock. The seismic method was useful in distinguishing between two formations which, in some areas, had about the same resistivities.

McGinnis and Kempton (1961) used a combination of seismic refraction, resistivity and geological methods in a study of glacial deposits in three areas in Illinois. Resistivity was useful for identifying and estimating depths to low-velocity sands and gravels, and for determining depths to the base of high-velocity layers in the drift. Both these resistivity results were exploited in the analysis of the seismic refraction data with a consequent improvement in the accuracy of the bedrock depth determinations. McGinnis and Kempton concluded that the joint use of the two geophysical methods in conjunction with the geological results eliminated most of the problems associated with the individual use of each.

Integration in western Canada. Reported western Canadian instances of combined geophysical surveys similar to those cited for other areas have been rare. However several consulting firms have some surveys on file and both federal and provincial government agencies are currently active in the field. The Geological Survey of Canada is co-operating with the Inland Waters Branch of the Canada Department of Energy, Mines and Resources in integrated studies in the areas of Winkler, Manitoba; Oak Lake, Manitoba; and Steelman-Frobisher, Saskatchewan. They are also establishing a new test area in the Manitoba section of the Virden map sheet (NTS 62F). Methods under investigation include dc resistivity, seismic, gravity and the airborne INPUT electromagnetic method. Test drilling forms an integral part of each investigation. The Saskatchewan Research Council is concentrating on the surface resistivity method and its correlation with test-drilling and electric logging results. This investigation is being conducted generally throughout the southern part of the province but there are some areas of concentrated study in the vicinity of Saskatoon. The Research Council of Alberta is continuing with resistivity, gravity and seismic studies in various parts of the province. Those combined surveys that have been made were initiated primarily for the comparison of the effectiveness of the various geophysical methods and not for the express purpose of either reducing geophysical exploration costs or of improving data quantity or quality.

Results of both individual and combined surveys to date suggest that it is certainly possible to improve data quantity, at least in some areas, by a judicious combination of geophysical methods. The principal example of such an application would be the combined use of gravity and resistivity exploration for tracing buried-valley courses. Similarly, data quality can be improved by utilizing resistivity or gravity to locate a buried-valley course and seismic refraction to estimate its dimensions. Seismic data quality was enhanced in the Two Hills area by supplementing the refraction measurements with borehole velocity measurements. However McGinnis and Kempton (1961) handled a similar problem in Illinois by using resistivity information, and this type of approach may also be possible in at least some areas of western Canada. Because it may not be possible everywhere, borehole velocity measurements should be made routinely in conjunction with all seismic refraction surveys. A similar conclusion holds for electric logging in resistivity exploration and for sampling and density measurements in gravity exploration.

A deplorable lack of pertinent data concerning the physical properties of near-surface deposits has been a major handicap in groundwater exploration by geophysical methods in western Canada. If such data were generally available it would be possible to predict which geophysical method or combination of methods should give the most satisfactory results in any given area and exploration efficiency should increase. Inclusion of sample measurements and logging techniques as standard operating procedures in shallow geophysical field studies is a necessary step in building up an adequate catalogue of physical data and the build-up process can be accelerated if groundwater geologists engaged on stratigraphic drilling programs will co-operate by providing core samples for physical measurements.

Representative costs for the three methods discussed in this paper vary considerably. Resistivity is the cheapest, particularly if the apparent-resistivity data are used simply for the derivation of isoresistivity maps and not for the determination of layer depths and thickness, an application for which, in any case, they are not particularly suited. Gravity is more expensive because of the necessity to measure station elevations accurately and because of the considerable effort involved in data reduction. Seismic refraction is more expensive still because of the need for relatively elaborate field procedures and because interpretation of the seismic data under the low velocity-contrast conditions prevalent throughout the Canadian prairies is generally far from a straightforward procedure. In the interests of keeping exploration costs down and maintaining exploration efficiency, it seems desirable to make as much use as possible of resistivity and gravity for the reconnaissance of large areas, and to save the seismic refraction method for those areas where detailed and fairly accurate results are required.

The exploration strategy outlined in the preceding paragraph assumes that all three methods are reasonably efficient in locating geological bodies or structures of interest, although the first two may be deficient in their ability to measure depths and thicknesses. It ignores the changes in strategy that may be necessitated by local conditions and it also ignores another and most important factor, the comparative costs of test drilling. Generally, experience in Alberta has shown that – on a purely economic basis and apart from any differences in the quality of the results obtained – resistivity and gravity exploration are competitive with test drilling; whether or not they are employed in groundwater exploration should depend therefore solely on the chances of obtaining useful information. Refraction exploration to depths of several tens of feet using hammer seismic equipment has a similar justification but refraction exploration to depths of less than about 150 or 200 feet using conventional equipment and explosive energy sources is, economically speaking, a rather dubious venture. For these shallow depths, the field and office costs of obtaining seismic estimates of layer depths and thicknesses for a given spread commonly exceed the costs of drilling a test hole to the deepest layer at the same location. Under these circumstances, use of the seismic refraction method in groundwater exploration is justifiable only on a research basis. The information obtained may be of value if future developments in instrumentation and techniques should lead to a substantial reduction in seismic exploration costs.

Although much effort has been expended in western Canada on shallow exploration for groundwater, until recently very little has been directed toward comparison of the effectiveness of the various methods in a diversity of geological settings. This situation is now undergoing a radical change but unfortunately the results of most of these comparative studies are either incomplete or still await publication. Candid unbiased evaluation of their results is essential if groundwater geologists are to be convinced that geophysics has any significant role to play in the search for groundwater. Too often in the past geophysics has promised much and delivered little and it should occasion no great surprise that many geologists have developed a hearty and abiding skepticism toward its potential as a groundwater exploration tool, at least in the geological environment of the western Canadian plains. If they are to be persuaded otherwise, they must be presented with incontrovertible evidence that the geophysicist can preselect on a rational basis the method or methods most suited for exploration in a given area and that he can, furthermore, provide a realistic and honest appraisal of the chances of success using these methods. The geophysicist must be willing to admit that there may be areas unsuited for the geophysical approach and in which it can be conducted only on an experimental basis with little or no guarantee of useful results. The ability to do these things depends on the availability of a good selection of case-history studies and on the availability of comprehensive catalogues listing basic physical and geophysical properties for near-surface surficial and bedrock materials. A concerted effort to carry out these case-history studies and to obtain these basic data should be the primary aim of groundwater geophysics on the Canadian prairies. Only in this way can the groundwater exploration potential of each geophysical method be assessed and its proper role as a member of an integrated exploration team determined.

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Discussion on Integration of Geophysical Methods for Groundwater Exploration in the Prairie Provinces, Canada, by D.H. Lennox and V. Carlson.

D.H. Hall, University of Manitoba, Canada. Do you have some case history verified by drilling and how useful it is? I would think that the conditions sometimes vary from one area where it is better to have some seismic profiles done with drilling at the end and to another where drilling alone might be cheaper.

D.H. Lennox, Research Council of Alberta, Canada. The work that I reviewed in the paper was partly our own and partly the work done by Saskatchewan. I think it depends on the depth to which you have to prospect. In Alberta, in most areas, we are talking about very shallow bedrock depths. I don't have case histories to cite the costs in areas where seismic might be cheaper. The northwestern part of the province, such as the Peace River district, might be an area where bedrock is deeper and seismic method is more feasible.

Hall. Then you would consider the use of seismic in areas where the bedrock depths are greater?

Lennox. Yes. Depths of the order of 150 to 200 feet which I quoted in my paper.

L.V.A. Sendlein, Iowa State University, U.S.A. I think this is a function of how many stations you can do in a day. The more stations that you can do in a day makes your survey costs much less. We get somewhere in the neighbourhood of ten reversed profiles a day and penetrate to 300 feet easily. We find, to drill a hole, we would need it as a last resort. We tie our data into holes drilled in the area. We find that we can operate on $120 per day, which includes living expenses. The crew consists of three helpers and a geophysicist.

Lennox. This is comparable to a crew that we use. It is a much slower operation. I don't know what the exact differences are though. We use four cables in our operation.

Sendlein. In both these cases, do you find that you are handicapped in not knowing the lithology but just knowing the bedrock without the lithology? By drilling a hole you get the lithology on the way down, as well as the bedrock. With your refraction, you get only the

depth to bedrock. You still don't know your lithology.

Lennox. Yes, at a single point such as over a drill hole, you get higher quality information. It depends on how you want to spend your money.

S.H. Raza, Geological Survey of Pakistan. Do you obtain information or reflections from layers in the drift?

Lennox. In our case, the buried valley sands and gravels that we are talking about are usually relatively thin, therefore we cannot detect aquifers close to the surface by the seismic method.

W. Meneley, Saskatchewan Research Council, Canada. I think that some points brought up here can be referred back to some points that Mr. Shiftan made. What are you looking for here in the context of regional hydrology? Certainly in the western plains, the mere location of the aquifer is not the critical point. In the case here of the buried valley aquifers, the permeability of the aquifer and the actual position has very little bearing except on where you want to put the well. The limiting factor in the developing of the resources, the regional hydrology, is the vertical permeability of the materials over the aquifer. This is what we have to determine. This is where you need your very detailed regional stratigraphy, in the tills above and beside the aquifer. This is what we so often don't get from, for example, seismic. Perhaps we get a little bit more from the resistivity methods but I don't think that we are getting the sort of resolution that we need to define the sort or permeability between the argillaceous sediments which overlay and bound our aquifer.

Lennox. I think we shall always have to resort to drilling for this information. Nonetheless, I think the function of geophysics is to help you find the proper location for the drill.

Meneley. I don't agree at all. Take, for example, the Estevan Valley aquifer. You can put wells into that aquifer, and one well can develop a pumping cone into Alberta and into the United States without any difficulty at all. It is a regional management and hydrogeological problem that you are dealing with in these major aquifers extending to a much greater extent than is actually realized. We are working in a very fortunate situation in Saskatchewan in that we have a very advanced technology in the investigation area prior to any major development in groundwater. We are essentially seeing virgin hydrogeological conditions in that we can see the tremendous extent of pumping cones. In a recent pump test, we were able to detect drawdowns 90 miles away on the second day of the pumping test at a few hundred gallons a minute. The cones overlap tremendously here. This puts a completely different scale on exploration. All groundwater developments interact and so you must deal with them on a regional basis.

Lennox. I feel this is a regional problem, all right, but somehow we have to locate the aquifers. I still think geophysics can help you in where to drill and where your observation points should be.

L.S. Collett, Geological Survey of Canada. I wonder if the hydrogeologists have defined this problem before to the geophysicists.

Sendlein. I get this from our own survey people all the time. I believe that geophysics has a very definite place in that it is a tool and one must think of it as a tool. A geologist must still be a geologist to find water and a geologist must understand the structures that he is working with. One gentleman said that we have to know vertical permeabilities to understand the system. In Iowa, the vertical permeability is not helping us because all our buried channels that we study are recharged from below. So we have not only got to understand the distribution of these aquifers at the surface but we also have to understand the subsurface geology.

Meneley. It is still the vertical permeability of the argillaceous sediments.

Sendlein. The point is that going through a detailed program as you have done (and I think it is an excellent program and one that can be considered optimum) costs a lot of money. What we are looking for is a way to do a program for a lower cost.

Meneley. I don't agree with doing a program for less money than you need. We need a lot more money in this business.

T. Hurr, U.S. Geological Survey. It depends partly on what kind of condition you have regionally. If you have uniform conditions such that you can put drill holes 5 miles apart, then you can use geophysics to give you more detail in between. But if you have extreme variations within a quarter of a mile, then you are going to have to do more drilling rather than being able to extend geophysics. You can only go between two known points. It all depends on the uniformity in whatever system you are working and I think that goes with what Prof. Sendlein was saying, too.

Hall. In light of what Dr. Meneley has raised, does your review of geophysics indicate any method which might be sensitive to permeability? I don't mean just methods or experiments. Maybe this is what we should be looking for. If water is moving, I suppose there is a self potential.

Lennox. Yes, but no correlation though.

Hall. Could an experiment be done whereby you can make water move.

Lennox. This would require the drilling of wells, again.

D. Gendzwill, Saskatchewan Research Council, Canada. We have been using geophysics in Saskatchewan. In some cases it seems to work fairly well and in other cases, it doesn't work too well. One of the most effective methods that we have been able to find for use in groundwater problems in Saskatchewan is well logging. A well-logging tool operates in a controlled environment. You can obtain very specific measurements on properties in drill holes. The technique holds considerable promise in exploitation study of resources, both for the study of aquifers and for the study of permeability that Dr. Meneley has mentioned.

Meneley. Borehole geophysics has been extremely useful. The problems that Scott Keys talks about in his paper offer hope in dealing with this problem of vertical permeability. The subsurface methods have proven successful and are offering further benefits in the future. The surface methods are marginal at best and, looking back, I would say that there isn't really any prospect that the surface

methods are going to help us with the problems that we have.

Lennox. I will go along with the fact that the surface methods are marginal. Nevertheless, we should continue these investigations as long as they are economical over other techniques.

Raza. Surface methods have certainly aided the choosing of drill well sites in Pakistan especially in the hilly regions. A well costs 10,000 rupees. Where there is no basis for drilling, no one likes to invest his money. Surface measurements do have a place in the art of groundwater exploration.

Meneley. In the context of regional hydrogeological studies, if you have done your regional geologic 'homework', then there is no problem at all in deciding where to put a well and deciding how deep it should go. In other words, you should have defined your base of exploration on a regional scale long before you get into detailed problems of shall we put a well here or 1/4 mile away.

Raza. That may be true in alluvial plains but what about mountainous valleys where the accumulation of sediments is very erratic. There you cannot put your guesses to work. You have to be very sure of your choice of drilling sites. In some mountainous valleys, the gravels extend from the base of the mountains in tongue shape and there you have to pinpoint those tongues. Therefore, I think geophysics is an important tool for determining the geology.

C.E. Anderson, Underwood, McLellan and Assoc., Canada. Maybe I can make the geophysicist feel a little bit happier. I am a geologist in the development consultant business in Manitoba. We have developed several water supplies from buried channels. For our purposes, to define the channel which is the only thing with which we are concerned, we have found that geophysical results are about two thirds of the cost compared to broad regional drilling.

Geophysical prospecting for groundwater in the Soviet Union

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Moscow State University

Abstract. Geophysics has been applied to groundwater exploration in the Soviet Union since 1929. The principal groundwater problems dealt with are: investigations of permeability and water-bearing properties of rocks; search for groundwater accumulations; and studies of groundwater movement.

Electrical prospecting, electrical logging, and to a lesser extent, seismic prospecting are the principal methods used to solve these problems.

Résumé. Les méthodes géophysiques sont appliquées à l'étude des eaux souterraines depuis 1929, en Union Soviétique. Trois problèmes ont été principalement étudiés: la perméabilité et les propriétés de rétention des roches, la recherche des nappes aquifères et l'étude du mouvement des eaux.

Les méthodes électriques, le carottage électrique et dans une moindre mesure les méthodes sismiques ont surtout été utilisés.

The first attempts to use geophysical methods in groundwater studies were made in 1929. In the main, electrical prospecting, electrical logging and seismic prospecting have been successfully applied in solving various hydrogeological problems in a few thousand cases. Thermometric and nuclear techniques have been applied under certain conditions. Gravimetric exploration and magnetic prospecting have been conducted occasionally.

The principal hydrogeological problems solved by geophysical methods are estimation of rock permeability and rock storage capacity; location of groundwater bodies and groundwater level determination; and groundwater movement.

These problems arise in the search for water-supply sources, in mine drainage and geological engineering studies for large-scale construction.

Estimation of rock permeability and rock storage capacity

The degree of water saturation of rocks and their capacity to transmit or retain water is closely related to some parameters being studied by geophysical methods. This relation is schematically presented in the accompanying table.

It is beyond the scope of this report to analyze the materials on which the above relationships are based, giving their physical interpretation. Therefore, the author will deal only with the problem that invariably arises in studying hydrogeological wells, that is, determining which high-water-yield well sections require screens.

As a result of numerous investigations carried out in different regions, V.A. Ryapolova has established approximate values of formation factor, that is, the ratio of rock resistivity ρ_r and water resistivity ρ_w ($F = \rho_r/\rho_w$), when fairly high water yield may be obtained from an aquifer. It is evident that V.A. Ryapolova's values are approximate and must be refined for each study of the petrographic parameters of the rocks, the salinity of groundwater and numerous other natural factors characteristic of the particular geological-hydrogeological region. Rock resistivity is measured in wells by means of logging tools of various sizes that allow the so-called 'lateral probing' to be conducted. By interpreting the curves obtained from such probing, the influence of the drilling mud filling the borehole and the zone of porous rocks where drilling mud penetrates is excluded. Groundwater resistivity is determined from direct resistivity observations in wells or can be calculated if the chemical composition of the groundwater is known.

Measurements of self potential (U_{ps}) and radioactivity (I_γ) appear to be useful subsidiary methods in determining sections with high water yields. The values of these parameters increase

Parameters related to water saturation of rocks.

Physical Parameter being Studied	With Increase in Water Content		With Increase in Permeability	
	In Loose and Plastic Rocks	In Consolidated and Semi-consolidated Rocks	In Loose and Plastic Rocks	In Consolidated and Semi-consolidated Rocks
Resistivity (ρ)	Decreases	Decreases	Increases	Decreases
Diffusion-absorption activity (A_g)	Increases	Increases	Acquires negative values	Acquires negative values
Polarizability (induced potentials)	Increases at first, then decreases a little	Relation is not established	Depends on the clay content & therefore, permeability	Decreases
Dielectric permeability (ϵ)	Increases	Increases	Relation is not established	Relation is not established
Velocity of propagation of elastic waves (V)	Increases	Usually increases a little	Relation is complicated & poorly studied	Decreases a little
Density	Increases insignificantly	Increases insignificantly	Relation is not established	Relation is not established
Slowing down & absorption of neutrons	Increases	Increases	Relation is not established	Relation is not established
Intensity of natural radioactive radiation	Relation is not established	Relation is not established	Decreases	Relation is not established

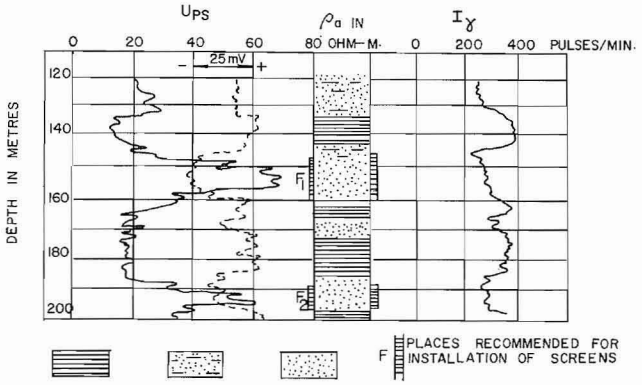


Figure 1. Determination of places for screen installation from apparent resistivity (ρ_a), self potential (U_{PS}) and radioactive (I_γ) logging data.

directly as the content of clay fractions increases, which in turn reduces the rock permeability and capacity to yield water. Thus, within porous beds and fractured zones, the well sections that correspond to the deepest parts of the minima of the diagrams U_{ps} and I_γ, are usually characterized by the highest specific water yield. Figure 1 gives an example of combined interpretation of the diagrams ρ_a, U_{ps} and I_γ in selecting sites for screens in clayey sand and carbonate rocks.

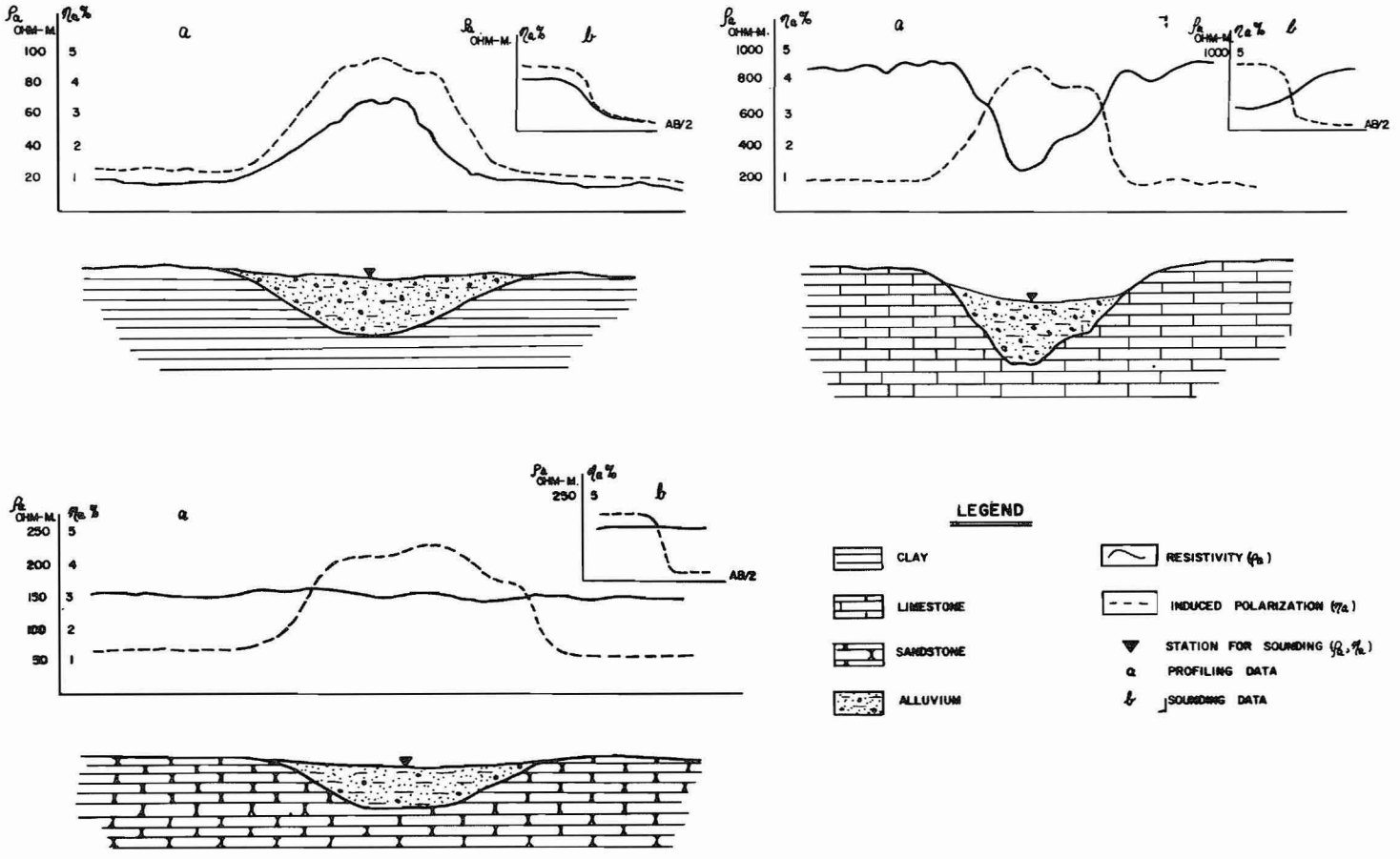
The results of the studies of hydrogeological sections of boreholes may be much more reliable if observations of rock response to neutron irradiation are carried out and ultrasonic logging is applied. However, these widely used methods in oil well studies have not been practically applied to hydrogeological problems.

Search for groundwater aquifers and groundwater level determination

Groundwater formation, occurrence and movement depend on lithologic-petrographic, geostructural, neotectonic, geomorphologic and other factors. The geophysical methods that may be applied in evaluating these factors and related problems are very diverse. Therefore, the author will describe only the major geological problems of groundwater investigations and their solution in general.

Much groundwater is taken from alluvial deposits. Therefore, geophysicists very often must explore for alluvial deposits and particularly those high in coarse detrital material, and estimate their depth and thickness. Such problems may, as a rule, be successfully solved by methods of electric, gravimetric and seismic prospecting. These methods have proved to be most effective in tracing old buried stream channels in the arid zones of Kazakhstan. Groundwater from these channels became the principal source of water supply for many settlements and

Figure 2. Typical apparent resistivity (ρ_a) and induced polarization (η_a) graphs for buried valleys.



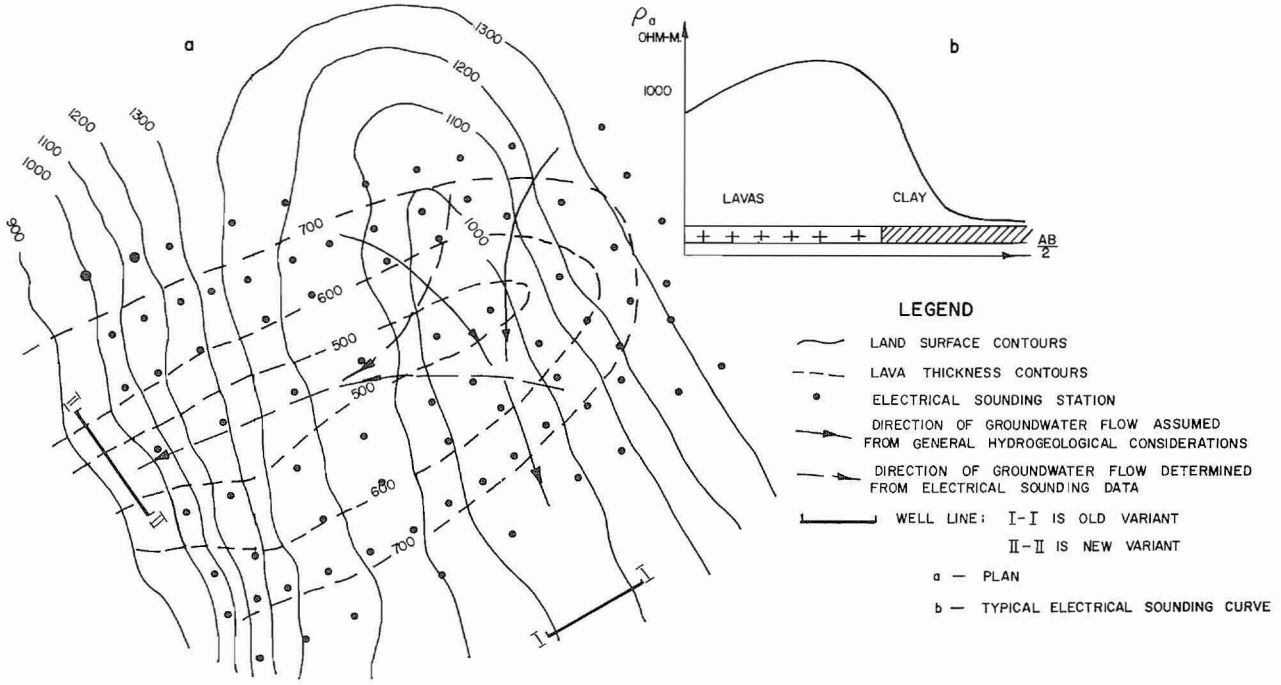


Figure 3. Comparison of land surface contours and lava thickness contours as determined from electrical soundings.

industrial enterprises. It should be pointed out that buried stream channels are commonly not exposed and numerous borings at close intervals are required to delineate them.

Figure 2 gives diagrammatic sections of some typical old buried stream valleys and graphs of geoelectric observation results. The character of the graphs depends primarily on the lithology of the bedrock in which the old stream channel was carved. For instance, alluvium within a buried clay bedrock channel is characterized by higher resistivity values, and within a carbonate bedrock channel by minimum values. However, there are cases where there are no differences between the resistivity values of various bedrock types. Under these conditions alluvial bodies may be detected by using the method of induced polarization, since fluvial deposits may differ from adjacent rocks in their higher polarization values η. Profiling and probing graphs for the parameter η are presented in the same figure.

High-accuracy gravimetric exploration is being successfully applied in the search for old buried stream channels, deeply incised in bedrock. Geophysicists obtain desirable results more easily and quickly by this method than by electrical prospecting. In many cases seismic prospecting is hindered by complications that are due to velocity inversion in the alluvial deposits. However under favourable seismogeological conditions seismic prospecting yields excellent results in determining the depth to the bedrock.

Investigators are very often interested in the configuration of the top of the impervious beds, since the direction of groundwater movement is primarily controlled by this surface and therefore wells can be drilled to obtain groundwater most effectively. Since impervious beds are mainly composed of clayey rocks that have low resistivity, the purpose of electrical methods is to determine the configuration of the interface between the pervious rocks, which exhibit a higher electrical resistivity, and the impervious beds. If the section of overlying beds is not too complex and the thickness of the impervious clayey substratum is considerable, this problem may be solved successfully by electrical probing using both direct and alternating currents.

Such investigations were conducted in one of the Transcaucasus regions, where thick strata of young lava flows with an average resistivity of 500 to 1500 ohm-metres occur over an area of tens of thousands of square kilometres. These lavas cover clays of Tertiary age which are characterized by a low resistivity of about 10 ohm-metres.

Rainfall readily percolates through fractured lavas and accumulates above the clay surface whence it moves to the low regions of sublava relief. Electrical probing with a maximum current electrode spacing of 10 km has in some areas provided information on the depth to impervious rocks over an area of about 4 thousand square kilometres. The calculated lava thicknesses ranged from some tens of metres to thousands of metres and in some localities the calculations were confirmed later by drilling. Figure 3 gives an idea of the behaviour of the sublava relief in one of the regions examined, where shortage of water was very acute. On comparing the contours of the sublava relief and the land surface, hydrologists were very surprised to find that the main groundwater flow must cross the watershed divide. It was suggested that wells drilled along line II - II which intersects a valley in the sublava relief discovered by electric probing would be more favourable for groundwater flow catchment than wells along line I - I that had been planned on the basis of general predictions.

Electrical prospecting has given excellent results in studies of artesian basins in which the structure largely resembles a syncline or syneclise. The best results are recorded on the areas corresponding to basin limbs where groundwater heads are mainly formed.

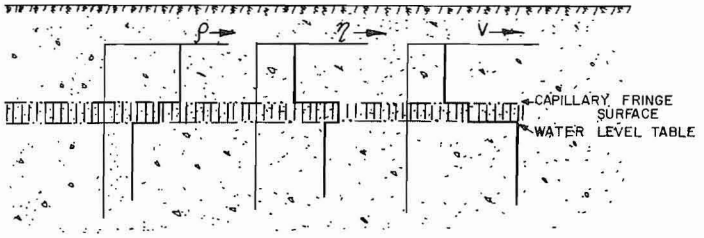


Figure 4. Change of electrical resistivity ρ, polarizability η and longitudinal seismic waves V at the boundary of zones of aeration and full saturation in gravel and sand strata.

On the basis of electrical probing data obtained from a fairly dense network, the investigator is able not only to determine the general character of the aquifer plunge, but also to distinguish separate positive structural elements. These positive structures in the productive layer are usually the most favourable sites for installation of wells, as the water yield of rocks increases because of high jointing and porosity in the arch part of the structures.

Toward the central parts of artesian basins where the rock thickens and the water is stagnant, geoelectric conditions become less favourable. The increased thickness of the rocks overlying the aquifer demands more elaborate and technically complicated deep probing with attendant difficulties in the interpretation of the results. On the other hand, the increase in the dissolved salt content of the stagnant water leads to a substantial decrease in the electrical resistance of the productive layer, which in turn reduces the resistivity contrast of the interfaces studied by probing. In soluble rocks, electrical conductivity inversion may occur and, as a result, the resistance of the aquifer becomes lower than that of the confining impervious top-layer. Under such conditions, gravimetric or seismic prospecting, the latter being the most accurate geophysical method, is preferred to electric prospecting.

The geophysical methods for determining groundwater level are based on the fact that certain physical properties of rocks change considerably depending on their water content (see table). Thus, a physical boundary is created between the undersaturated zone of aeration and the saturated zone below it. The behaviour of this boundary may be studied by making electrometric, seismometric and gravimetric observations.

In gravel and sand deposits, the level of the water-table is indicated by two electrical parameters, resistance and polarizability, as well as by a sharp change in elastic properties (Figure 4).

Induced polarization curves are characteristically more representative than ρ_a curves and provide the basis for a more reliable indication of the water-saturated surface. At the same time, one should bear in mind that induced polarization probing determines not the water-table proper but the depth to the capillary fringe. Induced polarization exhibits its maximum response within the zone of the capillary rise. This fact should be considered especially in conducting studies in fine-grained formations characterized by a large capillary fringe.

Unlike induced polarization and resistivity probing that give the depth to groundwater at certain points, the seismic refraction method is used to trace continuously the groundwater level, thus considerably increasing the accuracy of the results. In solving this problem, both the multichannel and single-channel refraction seismographs are usually used. Elastic waves are commonly generated by hitting the soil with a sledge hammer or by dropping a heavy weight from a special support.

The propagation velocities of elastic waves diminish considerably as the clay content of the rocks increases. At the same time, the induced polarization parameter increases with increase in the clay content up to about 10 per cent, then the polarization begins to diminish at the boundary of the zone of aeration and full saturation.

In loams and clays, electrical methods are practically useless for the determination of the depth to groundwater. Seismic methods have been used to some advantage though the results are less reliable than for sands.

In consolidated rocks, groundwater fills fractures and other interstices. Thus, apparent differences in the physical properties of consolidated rocks become more marked as rock porosity increases. Clay material that clogs interstices sharply decreases the ability to detect these differences. Of the three above-mentioned geophysical methods, only the resistivity probing gives fairly reliable results under favourable conditions. The differences in the seismic properties of consolidated rocks which vary in moisture content proves to be insignificant. The method of induced polarization holds little promise when applied to consolidated rocks, because of the generally low-level polarization (η) values in these rocks.

Hydrogeologists recently became interested in the possibility of using geophysical methods to study the behaviour of the groundwater table within cones of depression formed by pumping from mines and wells. As a rule, the number of observation wells is limited, resulting in less precise hydrogeological calculations based on the behaviour of the groundwater level within the cone of depression. Electrical probing and the more reliable refraction seismic method permit the tracing of the groundwater level in any predetermined direction and provide data concerning the peculiarities of cones of depression. Repeated observations in certain directions made during prolonged pumping permit the study of the variation with time in the size and shape of cones of depression. The movement of shallow groundwater with a low salinity to the place of pumping produces electrofiltration potentials which may be easily measured on surface. Figure 5 shows the results of observations carried out within the area of development of a stationary cone of depression formed by heavy pumping from a well drilled in South Armenia. The interference radius of pumping can be estimated from the maximum of natural electrofiltration potentials. Seismic observations along radii gave a continuous refracting boundary that coincides well with the water-table, the depth to which was determined by piezometric observations.

It is worth noting that accurate gravimetric observations permit the investigators to record the fluctuations of the density boundary that exists between the zone of aeration and the zone of full saturation as indicated in the table. Observations of this type were conducted in one of the large mineral deposits that occur within strongly fractured cavernous limestones in the northern Urals. Here, after a long period of heavy pumping from numerous mines, the groundwater level was lowered by more than 150 m. Water draining into mines or large karst cavities resulted in groundwater level fluctuations of tens of metres. Stationary measurements of gravitational forces permitted these

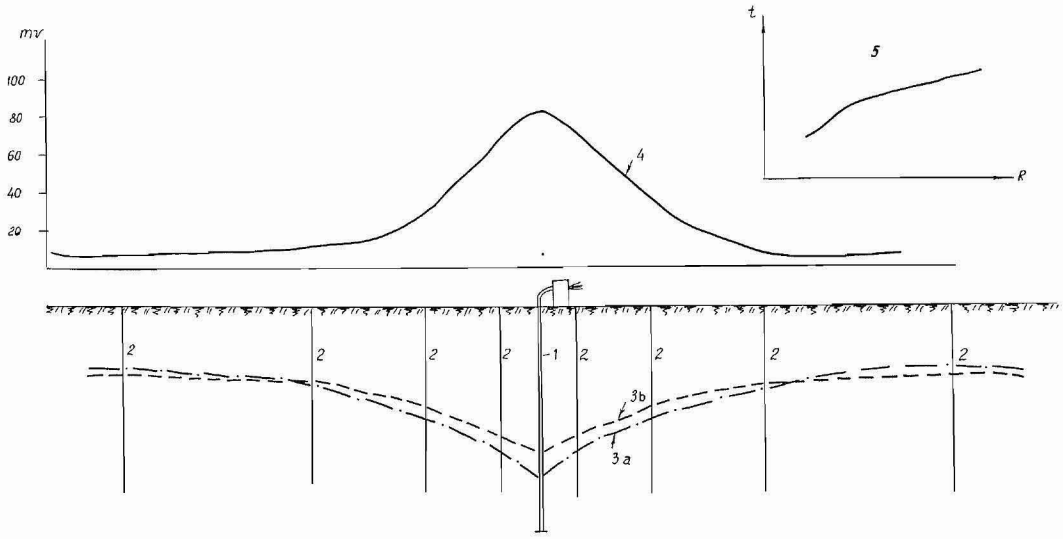


Figure 5. Determination of parameters of cone of depression from observation results using natural electric current and seismic refraction method. 1 – pumping well; 2 – observation well; 3 – water-table from data obtained in wells (a) from seismic prospecting data (b); 4 – curve of natural electric-filtration potentials; and 5 – refracted waves hodograph.

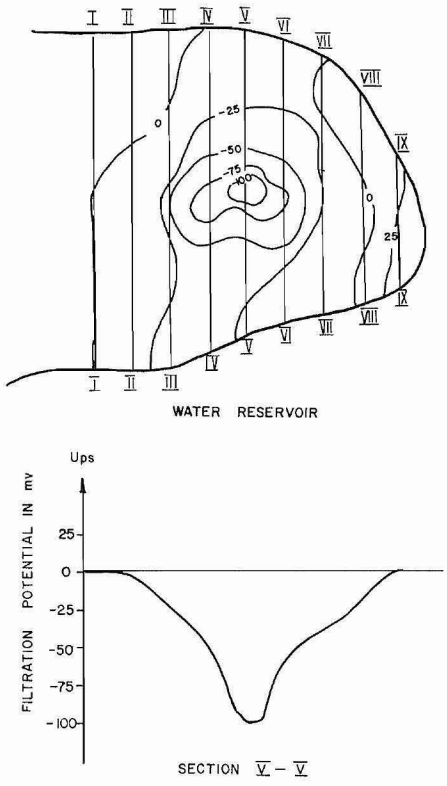


Figure 6. Anomalies of natural potentials above leakage zone in water reservoir.

fluctuations to be recorded reliably. These measurements became a useful addition to other observations carried out in this mining area by the Hydrogeological Survey.

Studies of underground water movement

In recent years, Grinbaum (1965), A.A. Ogilvy (1962) and N.A. Ogilvy and Fedorovich (1966) reported on the possibilities of using geophysical methods to study underground water movement.

The main objectives of these studies are determination of leakage sites in rivers, lakes and reservoirs; determination of areas of groundwater discharge under bodies of surface water or unconsolidated Quaternary deposits; tracing of shallow groundwater flows; determination of sites where water flows towards wells and estimation of filtration velocities under favourable conditions; determination of the vertical velocities of water movement in boreholes; and determination of the direction and actual velocity of groundwater movement in separate wells.

The sites of leakage from rivers, lakes and reservoirs are determined by the natural electric current method (Figure 6). Negative electric potentials appearing as a result of water movement downwards through the fractures or pores of rocks or loose soil on the bottom sometimes amount to hundreds of millivolts. The accuracy of the results obtained depends on the features and peculiarities of the water medium where the measurements are taken. Rivers, lakes and reservoirs have been examined using natural electric currents under different geological conditions. The most significant results are obtained in karst regions where considerable quantities of water are lost in sinkholes. Electric anomalies are significant where broad fractures and caverns are filled with medium-grained sand.

The measurements of natural potentials in reservoirs and lakes are supplemented by the study of temperature distribution in near-bottom water layers. In the warm seasons of the year, surface-heated water moves to leakage sites and thus positive temperature anomalies are created. After the leakage site is located, the investigators estimate water velocities in the direction of the leakage site. The velocity is obtained by measuring the resistance of heated wires placed both on the bottom and at certain levels above the leakage site. The temperature, and

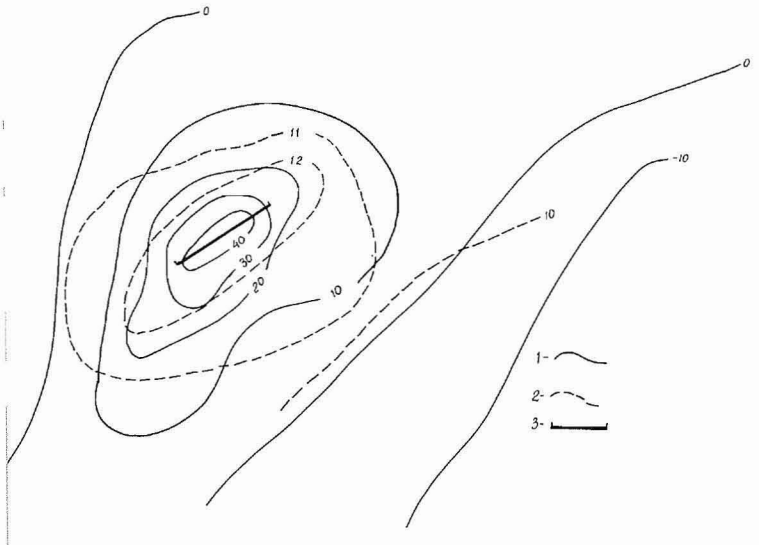


Figure 7. Anomalies of natural potentials and temperatures above groundwater discharge zone under sediments. 1 — isolines of natural potentials; 2 — isolines of temperatures measured in small wells; and 3 — assumed position of fractures through which groundwater discharges from bedrock (fractures are covered with sediments).

therefore the resistance of the wire change depending on the amount of water passing it. The apparatus is calibrated in advance under laboratory conditions thus allowing direct estimate of water velocities.

A detailed survey of apparent resistivities and natural potentials conducted along reservoir and lake shores permits the direction of water movement in rocks to be traced and correlated with leakage sites. Thus, by this method, leakage zones were found in one of the reservoirs in Armenia, where it was necessary to control the outflow by antifiltration screens.

Places of groundwater discharge into the bottoms of rivers, lakes and reservoirs may be determined by techniques similar to those applied in leakage studies. One should bear in mind that in such cases the inversion of the sign of the anomalies of filtration potentials will not be negative, as in the case of leakage, but will be positive. The water resistivity also increases. Water resistivity observations are particularly significant in determining the sites of fresh groundwater discharge into the sea bottom. Higher values correspond to these sites in resistivity diagrams.

If groundwater discharges in areas underlying Quaternary deposits, the areas of discharge may be found using data on the temperature, salinity and the amount of outflowing water. The thickness of the overlying unconsolidated deposits is also of great importance. In the search for zones of water discharge under Quaternary deposits, the natural electric current method and temperature measurements in small hand-bored wells are applied (Figure 7). A useful addition is a resistivity survey carried out by small electrical profiling equipment.

When shallow horizontal or near-horizontal groundwater flows are to be traced, the natural electric current method is employed using a radial-type electrode configuration. Filtration potentials are a maximum in the direction of the flow. In this connection, the radial diagrams constructed for observation points above the flow will be, because of their elongated form, considerably different from the character of the data measured at a distance from the flow. If water moves through fractures, the distribution of the latter may be shown by circular resistivity contours constructed from detailed profiling mainly using asymmetrical sondes.

Resistivity measurements and thermometry have been applied for some time to determine zones where groundwater flows to boreholes. Reliable results may be obtained only when the salinity and temperature of inflowing water differ considerably from those of the water or drilling mud filling the borehole. Therefore the water salinity in the borehole should be changed artificially and sometimes the water temperature should be increased or decreased.

The resistivity method is widely used in studying groundwater flow filtration velocities when the flow intersects the borehole. In the well under investigation, logging is carried out using the resistivity and self-potential methods to determine the general electrical characteristic of the geological section and to determine the approximate location of the aquifer. Then, along the length of the borehole the water resistivity is determined by a fluid resistivity meter; in the case of sufficient differentiation in the log, the places where water having a different salinity flows in are distinguishable.

After this preliminary procedure, a highly saline electrolyte is introduced in the well by repeatedly lowering sacks of table salt into the well. The salinization of the well is completed when the electrolyte concentration reaches 2 to 3 grams per litre, corresponding to a resistivity of about 2.5 ohms. The uniformity of the distribution of the electrolyte concentration is shown by a control resistivity log made along the length of the borehole or within the zone subjected to investigation.

When the electrolyte has been evenly distributed, water resistivity measurements are repeated, i.e., resistivity logs are made to determine the velocities of electrolyte washing away from the borehole. The time intervals between records range, depending on the quantity of desalinizing water, from 15 to 30 minutes, to several days or even weeks. The time at the beginning and completion of the water-resistivity log is recorded.

A series of such resistivity diagrams gives a picture of the relative rate of electrolyte washing away in various sections of a well (Figure 8). The places where electrolyte is desalinized most rapidly, i.e., places where the diagrams show considerable shifting with time, approaching the log of liquid resistivity under natural conditions, will obviously correspond to sites with intensive groundwater inflow. The quantitative interpretation of the zone within which resistivity diagrams are regularly shifted may be undertaken if 3 to 5 diagrams are obtained.

A number of investigators have established the law of change in electrolyte concentration in a well as affected by groundwater flow. This low may be expressed by the formula

$$W = \frac{1.81d}{a(t_n - t_1)} \lg \frac{C_1 - C_0}{C_n - C_0}$$

where W is the filtration velocity in the rock; d is the well diameter that is determined using a caliper meter; C_0 is the natural electrolyte concentration in groundwater; C_1 and C_n are electrolyte concentrations in the well at the time moments t_1 and t_n; a is the coefficient of the well wall permeability varying from

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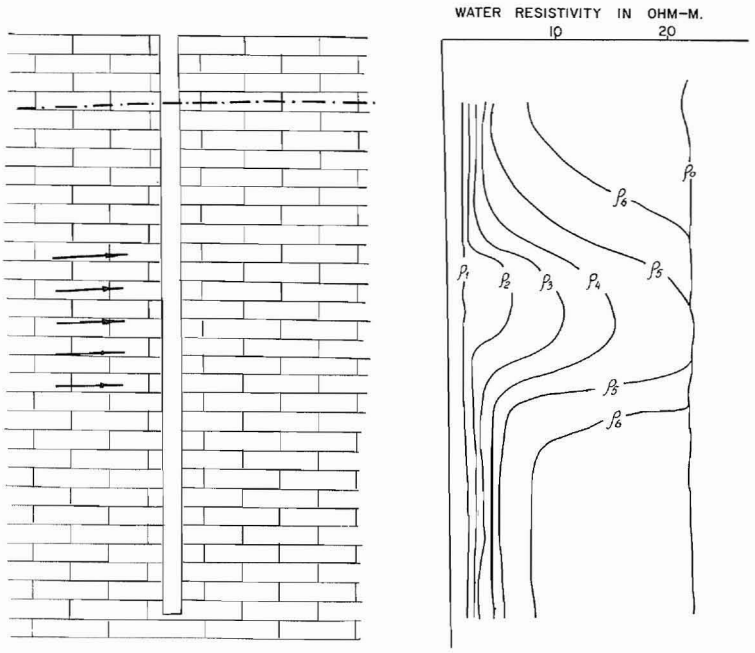


Figure 8. Determination of filtration velocity using resistivity method. Diagrams of water resistivity: ρ_0 — under natural conditions; ρ_1 — immediately after salinization; ρ_2, ρ_3, ρ_4, ρ_5, ρ_6 — records every 2 hours; —.—.—.— water-table; and → direction of water movement in zone of highest permeability.

0.5 to 4 and 2 for an uncased well with unmudded off-walls.

Electrolyte investigations to determine groundwater filtration velocities may be conducted both in uncased and cased wells with screens. In the latter case, it is necessary to make a correction for screen resistivity.

Testing of many hydrogeological and reconnaissance wells gives very valuable practical results. Even in karst regions with their highly complicated hydrogeological conditions, general hydrogeological concepts can be established by the water-resistivity method.

For natural flowing water or its leakage from one aquifer to another, flow meters are applied allowing specific discharges to be estimated at various sections in the borehole. The flow meter employs either the principle of heated wire resistance that depends on the velocity of water movement in the borehole or light rotators. Such observations give data on the yield of aquifers and the water which flows to the well.

The direction and actual velocity of groundwater movement in separate wells are determined by observations of the deformation of artificial electric fields due to electrolyte displacement in water flow. Electrolyte is introduced into the flow through an uncased or screened well. A current electrode A is lowered to a depth h corresponding to the middle of the aquifer or screen. A second current electrode B is moved to infinity, i.e., at a distance at least 10 times greater than h. The resistance of electrodes A and B must be small so that it will be possible to send current of more than 0.3 to 0.5 ampere into the ground.

Potential fields using an electrode N are measured on radii extending in all directions from the borehole. Equipotentials are plotted.

The electrolyte introduced into the well gradually penetrates into the surrounding rocks. The penetration will be naturally deepest in the direction of groundwater movement. In a homogeneous filtrating aquifer, the volume saturated by the electrolyte will take the form of an elongated tongue. Thus, at a certain depth below the land surface, this volume of high electrical conductivity will increase with time. This body will distort the electric field that existed at the moment when the electrolyte was introduced into the well.

If the position of equipotential lines is recorded at certain periods of time, they will be found to be displaced with respect to the original position. They will be elongated in the direction of movement of the body of high electrical conductivity, that is, in the direction of groundwater flow. To obtain more reliable data, several equipotential lines are recorded successively.

If groundwater moves at more than 3 m per day, these lines are recorded continuously. If the flow rate is lower, an interval between successive recordings should be 1 to 2 hours. All well observations usually continue from 8 to 10 hours, to 2 to 3 days. During this time, empty sacks which had contained the salt are removed from the well and new sacks of salt are lowered into the well. Thus the electrolyte is supplied to rocks uninterruptedly.

The most probable direction of groundwater movement is determined from the average maximum displacement of all equipotential lines. The actual flow velocity V is found as a ratio of the salt body displacement Δl and the period of time $t_2 - t_1$ during which this displacement took place:

$$V = \frac{\Delta l}{t_2 - t_1}$$

Satisfactory results can be obtained by the charged body method if the depth to the aquifer under investigation does not exceed 80 to 100 m. The average error in estimating groundwater velocities by hydrogeological and geophysical methods is about 25 per cent. In clayey rocks, groundwater velocities are sharply underestimated because of decrease of coefficients of permeability when concentrated electrolyte moves through clayey rock pores. Under these conditions, only the flow direction may be determined. If groundwater has a high saline content, this method cannot be applied.

In the case of recharging aquifers, electrolyte resistivity measurements may be also used to determine their absorbing capacity. In this case, the velocity of the shift of the boundary between electrolyte existing in the well and the fresh water injected at a constant rate along the borehole is recorded by a fluid resistivity meter.

There are also a number of other geophysical methods that may be applied to test pumping and injection.

Conclusion

Only the main general hydrogeological problems that may be successfully solved by geophysical methods have been analyzed. However the possibilities of application of geophysical methods are not limited to solving these problems. Geophysical methods are widely used in groundwater studies in permafrost regions covering about 43 per cent of the area of the Soviet Union. They are commonly applied in determining the interface between fresh and salt waters near sea shores and particularly in the search for

fresh-water lenses that are principal water-supply sources in deserts.

The report makes no claim to describe in detail the application of geophysical methods in groundwater studies.

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The use of seismic refraction and gravity methods in hydrogeological investigations

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Abstract. The capacity of unconsolidated deposits to store water depends on their porosity, thickness and shape; thus geophysical methods that provide information relating to these factors provide techniques for both search and evaluation.

Unconsolidated sediments commonly have low compressional wave velocities and low densities. Although many factors have a bearing on velocity and density, porosity is a particularly important factor. Low density and low velocity are a direct indication of high total porosity and potential water storage capacity, and hence a useful guide to unconsolidated aquifers.

Two types of deposits are considered: river valleys filled with fluvial sediments, stratified drift or glacial till; and intermontane basins filled with poorly consolidated fan gravels and sheet flood deposits. Under ideal circumstances, the seismic method provides a means of determining the three-dimensional geometry of the aquifer, the gross stratigraphy and local lithofacies variations of the aquifer, and depth to the water-table. The gravity method can be used alone as a reconnaissance tool for evaluating bedrock topography beneath the aquifer in some areas, and locally for quantitative estimates of depth to bedrock. Used with the seismic method or with borehole data, surface measurements of gravity provide a means of determining the average total porosity of the deposit and, in some instances, even a measure of specific yield.

Résumé. L'aptitude des sédiments meubles à emmagasiner de l'eau est fonction de leur porosité, leur épaisseur et leur forme; par conséquent, les méthodes géophysiques capables de fournir des renseignements sur ces caractéristiques offrent des techniques qui permettent d'en faire la prospection et l'évaluation.

Les sédiments meubles sont caractérisés en général par les faibles valeurs de leur vitesse de propagation des ondes de compression et de leur densité. Bien que ces deux dimensions soient fonctions de nombreux facteurs, elles dépendent essentiellement de la porosité. De faibles valeurs de densité et vitesse de propagation sont les indices directs d'une porosité d'ensemble élevée et d'un pouvoir d'emmagasinage important; elles permettent donc de rechercher les terrains aquifères friables.

Les terrains sédimentaires envisagés se divisent en deux catégories: les alluvions fluviales, les terrains de transport stratifiés ou les dépôts glaciaires; et les sédiments des bassins encaissés composés d'amas de gravier et dépôts d'inondation faiblement consolidés. Dans les conditions idéales, la méthode séismique permet de déterminer la géométrie à trois dimensions du terrain aquifère, les traits dominants de sa stratigraphie et les variations locales de son lithofaciès, ainsi que la profondeur du niveau hydrostatique. La méthode gravimétrique peut s'utiliser seule comme outil de reconnaissance pour déterminer la topographie de la roche en place sous les nappes aquifères en certaines régions, ainsi que, localement, pour procéder à l'évaluation quantitative de la profondeur de cette même roche. L'utilisation conjointe des mesures séismiques, ou de celles effectuées dans les forages, et des données fournies par la gravimétrie au sol permet de déterminer la porosité moyenne de la roche magasin et même, en certains cas, la valeur d'une nappe donnée.

Unconsolidated sediments as aquifers

The unconsolidated sedimentary fill of subaerial valleys is a source of abundant groundwater, whether deposited as fan gravel and sand by streams and sheet floods in interior basins, as fluvial sediments in, or adjacent to, present-day rivers, or as valley trains in watercourses that once headed at the margins of glaciers. The silt, sand and gravel members of such deposits generally constitute aquifers of high productivity and large storage capacity. Their capacity to store water depends on their porosity, thickness and shape, and their ability to transmit water depends on their pore texture. They are sought as aquifers because their porosity, specific yield and permeability are generally higher than those of other geological materials (with the possible exception of young, highly fractured lava flows and cavernous limestones) and because the development of wells in unconsolidated sediments is relatively simple and inexpensive.

Part of the information needed by the hydrogeologist attempting to evaluate an unconsolidated deposit as an aquifer is its thickness, horizontal extent, porosity and specific yield. Information of this kind can be obtained by a combination of geological mapping, stratigraphic analysis and the drilling of boreholes for sampling and *in situ* physical tests and measurements, but comprehensive studies of this kind are time-consuming and expensive. In some instances, they may not be warranted and surface geophysical methods can be used to economic advantage. Although the information the latter methods provide is not as specific nor as free from ambiguity as that obtained by drilling and borehole tests, it is often useful enough because of its relatively low cost to warrant serious consideration, particularly where it can be used to provide information that will help reduce the amount of subsequent test drilling or will allow extrapolation of available data where drilling has been done previously on a limited scale. Seismic refraction studies cost five to eight times as much as gravity studies, but both are considerably cheaper than drilling. According to Welin (1958), shallow seismic refraction measurements of bedrock depth in some areas are nearly 20 times cheaper than the drilling of boreholes. The choice of gravity or seismic methods in hydrogeology depends on the objectives of the investigation and on the geology of the area.

Table I. Compressional wave velocities observed in Quaternary and upper Tertiary unconsolidated clastic sediments (composited from published and unpublished data as individually noted).

Material	Minimum Vp		Maximum Vp	
	(km/sec)	(ft/sec)	(km/sec)	(ft/sec)
(A) Glacial and glaciofluvial sediments (data from Brown and Robertshaw, 1957; Clark, 1966; Johnson, 1954; Linehan, 1952; Linehan and Murphy, 1962; McGinnis and Kempton, 1961; Vecchia, 1958; Warrick and Winslow, 1960; and Welin, 1958)				
Stratified drift, including silt, sand and gravel	.35	1,150	1.86	6,100
Till and morainal debris	.43	1,400	3.00	9,850
(B) Alluvial fill of intermontane basins, expressed as composite velocity layers for seven basins in California and Arizona (data from Healy and Press, 1964; Kane and Pakiser, 1961; Mabey, 1956; Pakiser and Kane, 1962; Pakiser, Press, and Kane, 1960; Zbur, 1963, and unpublished data of the writers)*				
Layer				
V_0	.34	1,100	1.49	4,900
— water-table —				
V_1	1.46	4,800	2.26	7,400
V_2	1.71	5,600	3.29	10,800
V_3	2.16	7,100	3.29	10,800
V_4	2.44	8,000	3.90	12,800
V_5	3.14	10,300	4.11	13,500

*Some of the values in B, perhaps all those in excess of 3.05 (10,000 ft/sec), may represent upper Tertiary, semiconsolidated continental sediments.

Table II. Bulk densities observed in Quaternary unconsolidated clastic sediments (data from Hall and Hajnal, 1962; Manger, 1963; McGinnis, *et al.*, 1963; and unpublished data of the writers).

Material	Dry bulk density			Saturated bulk density		
	Min.	Max.	Approx. Avg.	Min.	Max.	Approx. Avg.
(A) Glacial and glaciofluvial sediments						
Stratified drift, including silt, sand and gravel	1.54	2.12	1.73	1.88	2.40	2.02
Till	1.82	2.19	2.07	2.08	2.36	2.26
(B) Alluvial fill of intermontane basins						
Undifferentiated clay, silt, sand and gravel	0.88	2.22	1.63	1.55	2.39	2.02

Table III. Total porosities observed in Quaternary unconsolidated clastic sediments of continental origin.

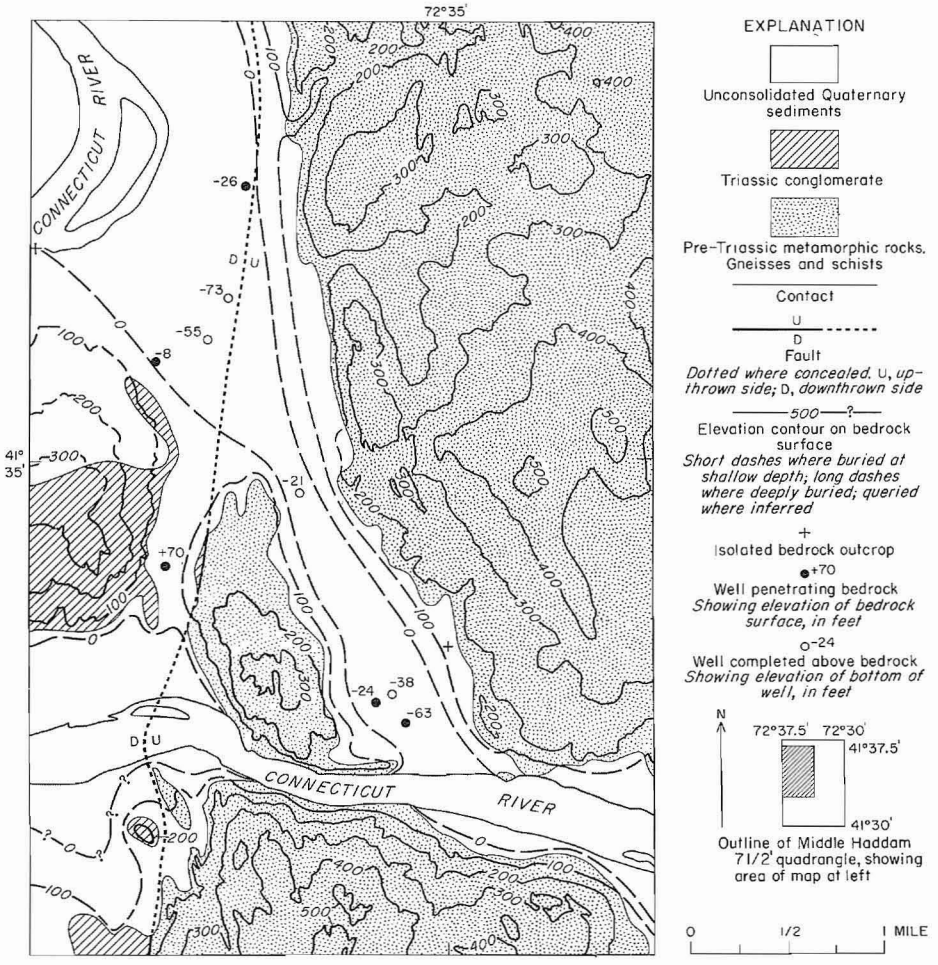
Description of material	Porosity		Reference
	Minimum	Maximum	
Sand, clay and soil	0.02*	0.48	Lee (1912)
Fine, sandy loam to coarse sand	0.39	0.54	Ellis and Lee (1919)
Clay and silt	0.38	0.43	Melcher (1921)**
Clay, silt, sand and gravel	0.18	0.48	Meinzer (1923)
Clay, silt, sand and gravel	0.23	0.53	Stearns (1927)
Sand and gravel	0.17	0.45	Eckis and Gross (1934)
Clay, sand and gravel	0.23	0.42	Wenzel (1936)
Silt, sand and gravel	0.33	0.60	Piper, *et al.* (1939)
Sand and gravel***	0.25	0.41	Jones and Buford (1951)
Clay, silt and sand	0.25	0.74†	Hall and Hajnal (1962)
Silt, sand and gravel	0.23	0.67	Cohen (1963)

*Lee's method of porosity measurement is not believed by the writers to have been accurate for fine-grained materials and the figure of 0.02 is probably neither correct nor representative. **As reported by Manger (1963). ***May include interbedded sediments of deltaic and near-shore, marine origin. †The maximum value of 0.74 reported by Hall and Hajnal represents a singular value on one porosity log. The *average* maximum value displayed by this log is approximately 0.58.

Seismic and gravity methods have been used widely to determine the thickness and areal extent of unconsolidated deposits and in the authors' view, under ideal circumstances they can also yield useful estimates of average total porosity. However, neither can provide information on quality of groundwater and the seismic method is of little help as yet in estimating permeability, though tests (Berson, *et al.*, 1959) have engendered hope that they may find qualitative applications to this end. The success of these methods in some applications depends largely on the existence of measurable contrasts in velocity and density between the fill and underlying bedrock and in the ability to translate these measured quantities into meaningful hydrogeological parameters (e.g., stratigraphic thickness, where velocity layers and stratigraphic units can be correlated; porosity; and to a first approximation, specific yield).

Many factors have a bearing on the velocity and density of unconsolidated sediments, but one of the most significant and quantitatively important is porosity (Hamilton, *et al.*, 1956; Wyllie, *et al.*, 1956; Nafe and Drake, 1957; and Sutton, *et al.*, 1957). Because relatively low density and low velocity are a direct indication of high total porosity and potential water-storage capacity in clastic sediments, they are useful in locating and evaluating unconsolidated aquifers. For this reason, and to provide some background data for discussions which follow, porosity measurements are included here with tabulations of velocity and density.

Some physical properties of unconsolidated deposits. One limitation of the seismic and gravity methods in studies of unconsolidated sediments is evident from reported values of compressional wave velocity, density and porosity (Tables I, II and III). If these physical properties, as they occur in specific deposits, do not contrast sufficiently with those of the bedrock beneath them, or if they vary greatly within the sediments themselves, many applications of the methods will fail.



1(a)

Figure 1. (a) Generalized geological map of a portion of the Middle Haddam quadrangle, Connecticut, showing distribution of outcrops of bedrock and elevation contours on the exposed and buried bedrock surface. Note buried channel of the preglacial Connecticut River. (b) Complete Bouguer gravity map of the same area.

Table IV. Range of compressional wave velocities and bulk densities to be expected in pre-Tertiary bedrock underlying younger unconsolidated sediments.

(A) Compressional wave velocity	(km/sec)	(ft/sec)
Minimum value	1.95	6,400*
Maximum value	7.32	24,000**

(B) Bulk density	Minimum (gm/cc)	Maximum
Dry bulk density	2.04	2.82**
Saturated bulk density	2.28***	2.82**

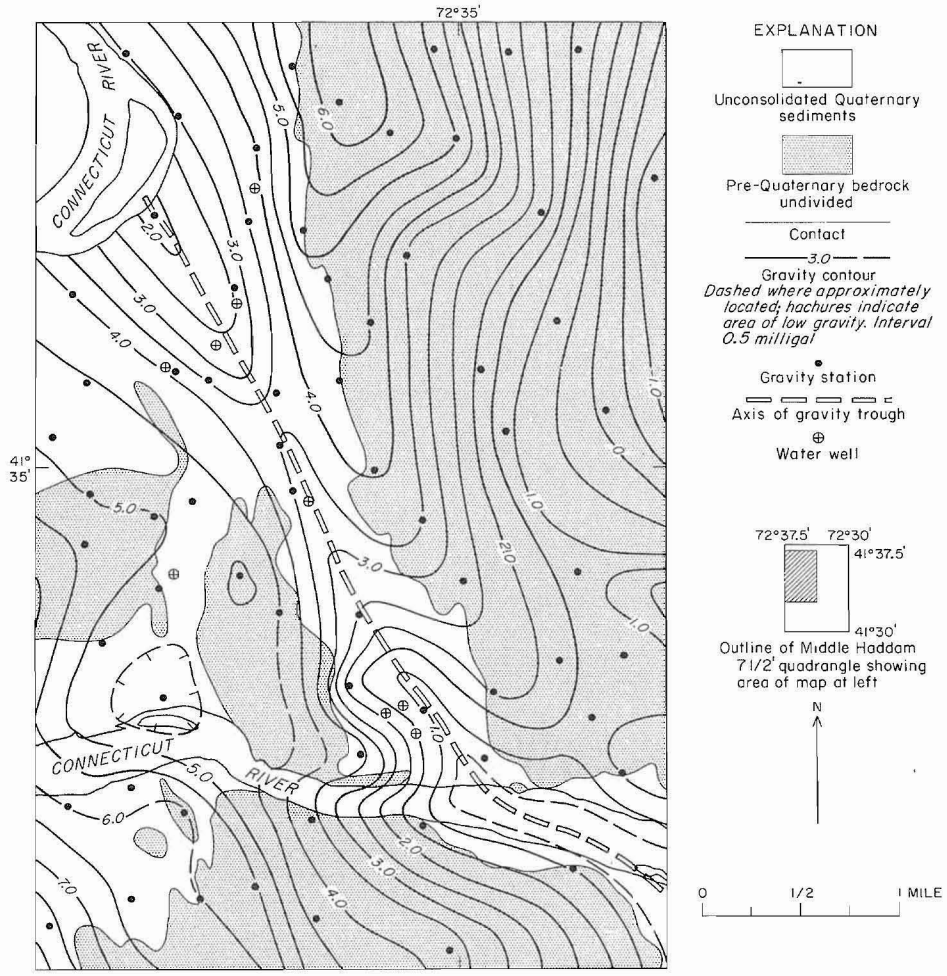
*Value of minimum horizontal compressional wave velocities for Upper Cretaceous sandstone and shale reported by Lennox and Carlson (1967, Table 2). **Unpublished value for Lower Paleozoic dolomite measured by one of the writers. ***Average bulk density for nonclayey, Cretaceous clastic rocks listed by Manger (1963, p. 18 - 21).

Variations in compressional wave velocities of unconsolidated or semiconsolidated sediments are the result of variations in many geologic factors, among which are depth of burial, lithology (including grain size distribution, grain shape, degree of cementation and composition of the cement), confining pressure, density, porosity, degree of saturation and fluid pressure of the saturant. Porosity itself is a function of many of these factors and for this reason, velocity and porosity display a very high degree of correlation. This fact is of considerable importance in hydrogeology.

A thick section of unconsolidated sediments, derived at a uniform rate from a single source terrain under essentially unvarying conditions, is fairly homogeneous in porosity and is relatively simpler to work with seismically than one in which the source materials, mechanisms of transport and deposition, and overburden pressures have varied with time.

Unconsolidated sediments of mixed glacial and glaciofluvial origin, especially those that have accumulated in areas of multiple glacial advance, typify the latter type and are known to cause considerable difficulty in seismic interpretation in some localities (see, for example, Lennox and Carlson, 1967, Figure 12).

Variations in the *in situ* bulk density of unconsolidated sediments are also the result of variations in many factors, but density can be completely defined by three independent variables—grain density, porosity and pore fluid density. Other



1(b)

variables have an influence on density to the extent that they give rise to variations in these three. As in seismic velocity, spatial variations in bulk density within an unconsolidated deposit may reflect variations in porosity, and at the same time they introduce difficulty to an interpretation of seismic data.

The total porosity of a granular aggregate (the percentage volume of the aggregate occupied by voids) depends on the shape, sorting and spatial arrangement of the grains constituting the aggregate, as well as on its geological history. Errors in the laboratory measurement of sediment porosity are appreciable under some circumstances. For unconsolidated, or weakly consolidated, sediments, these errors stem partly because collecting samples is difficult without physically disturbing the packing of the constituent grains. In addition, some unconsolidated sediments cannot be sampled at depth by drilling and information on their porosity is restricted to the upper part of the deposit, which may not be representative of the whole section. Because variations in porosity are indicated by variations in compressional wave velocity and bulk density, the geophysicist is, in favorable circumstances, in a position to aid the hydrogeologist to estimate porosities at depth in some sedimentary deposits.

Contrasting physical properties of older rocks. Because Tertiary continental sediments constitute productive aquifer systems in some regions, Upper Cretaceous porous sediments have been selected for Table IV as representing the minimum values of velocity and density that one might expect to encounter beneath post-Cretaceous sediments in seismic or gravity investigations of unconsolidated aquifers. The velocity and density of still older, nonporous rocks provide reasonable values for the maximum contrasts. The range of values in Table IV therefore indicates *approximate* limits for the average velocity and density to be found in rocks beneath most unconsolidated or semiconsolidated deposits.

A comparison of data in Tables I and II with those in Table IV indicates that glacial till and other compact, clay-rich, glacial sediments have compressional wave velocities that exceed those of some Cretaceous sedimentary rocks. Similarly, velocities of Cenozoic continental sediments of some intermontane basins may exceed those of Cretaceous sandstones deposited and lithified in grossly different environments. The same is true for contrasts in bulk density, for minimum values of Cretaceous sedimentary rocks fall *within* the minimum-maximum range indicated for Quaternary sediments of all types under both dry and saturated conditions.

Despite the potential difficulties suggested by a lack of contrast with bedrock on the one hand and by large intradeposit variations on the other, successful interpretations, confined by later drilling, or aided by concurrent drilling, have been made in

many areas. Bedrock values of velocity and density generally do tend to exceed those of younger unconsolidated or semiconsolidated sediments, and both properties tend to increase downward in the section.

Applications of seismic refraction and gravity methods in mapping unconsolidated deposits

Cited below are selected examples of seismic and gravity studies of unconsolidated deposits drawn from the more readily available recent literature in English.

Seismic refraction investigations of valleys filled with young unconsolidated sediments have been conducted by many investigators. In almost every study the objective has been a determination of the thickness or configuration of the valley fill.

Studies of large intermontane basins, primarily in the western United States, have been made by Biehler, *et al.* (1964), Diment, *et al.* (1960), Healy and Press (1964), Kane and Pakiser (1961), Kovach, *et al.* (1962), Mabey (1956), McGinnis and Dudley (1964), Pakiser, *et al.* (1960, 1964), Pakiser and Kane (1962, 1965), and Zbur (1963).

Seismic refraction studies of sediment-filled river valleys, buried or modern, have been reported by Bonini and Hickok (1958), Coster and Gerrard (1947), Dudley and McGinnis (1964), Hawkins (1961), Lennox and Carlson (1967), Mathiez and Astier (1958), McGinnis and Dudley (1964), McGinnis and Kempton (1961), Stam (1962), Vecchia (1958), Wantland (1959), Warrick and Winslow (1960), Watkins and Spieker (1965), Welin (1958), and Woollard and Hanson (1954). A similar study of river channels in ancient sedimentary rocks was made by Pakiser and Black (1957). Efforts to use reflection techniques in the study of unconsolidated sediments, some of which fill buried valleys, have been reported by Allen, *et al.* (1952) and by Pakiser and Warrick (1956).

Seismic studies of sheet-like deposits of young unconsolidated sediments, wherein the objective was either bedrock depth measurement or differentiation of stratigraphic units within the sediments, have been the subject of papers by Brown and Robertshaw (1957) and Johnson (1954).

Gravity studies of valley-fill deposits in intermontane basins have been made by Cook and Berg (1961), Cook, *et al.* (1964), Corbato (1963), Diment, *et al.* (1960), Garland, *et al.* (1961), Healy and Press (1964), Kane and Pakiser (1961), Mabey (1956, 1960, 1963, 1964), Pakiser (1961), Pakiser, *et al.* (1960, 1964), Pakiser and Kane (1962, 1965), Stewart (1958), Thompson (1959), Thompson and Sandberg (1958), and Von Huene (1960).

Gravity investigations of sediment-filled stream valleys, though less common than those of intermontane basins, have been reported by Cartwright, *et al.* (1964), Eaton, *et al.* (1964), Hall and Hajnal (1962), Heigold, *et al.* (1964), Lennox and Carlson (1967), McGinnis, *et al.* (1963), and Woollard and Hanson (1954).

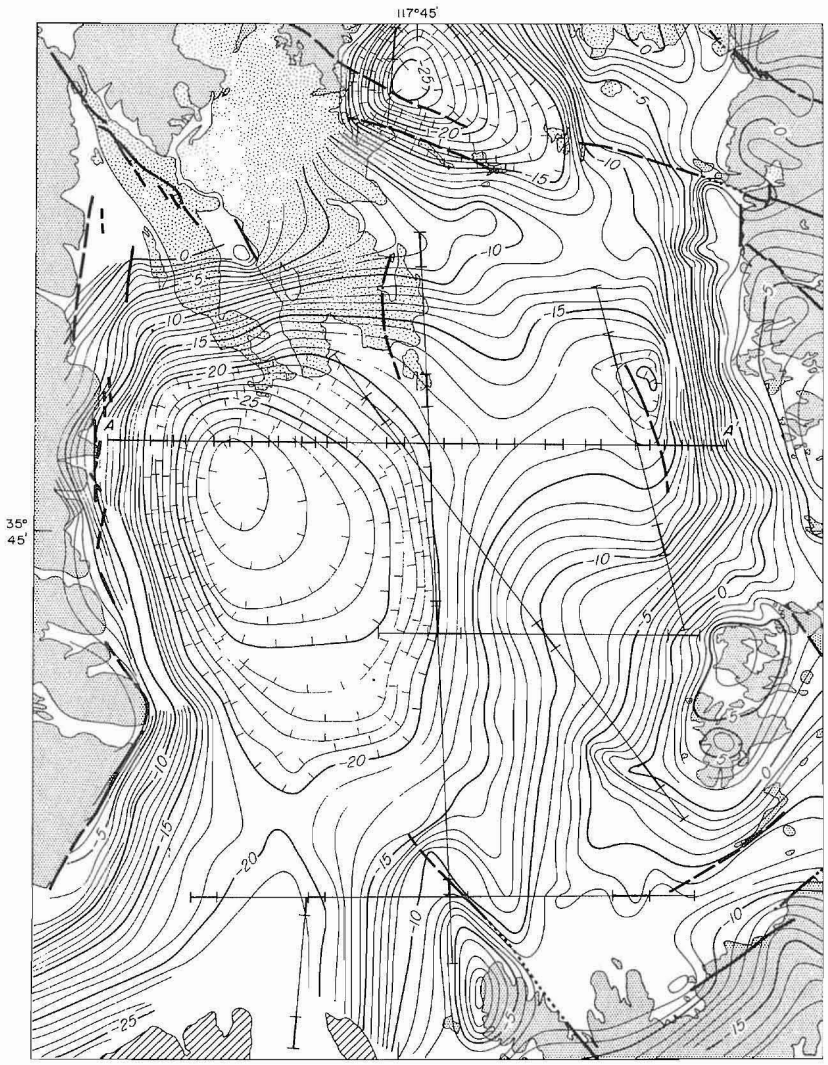
Brief sketches of representative case histories of each kind, including unpublished work by the authors, are presented to illustrate applications of the two methods as they pertain to hydrogeology. They are followed by a short discussion of some of the errors involved.

Gravity case histories

Gravity study of a buried river channel. The area depicted in Figures 1a and 1b is in central Connecticut. 1a shows the distribution of bedrock outcrops and the generalized geology of the area. The geological and gravity mapping was done by one of the writers in 1956. The area straddles a major normal fault separating Triassic sedimentary rocks on the west from Paleozoic metamorphic rocks on the east. Elevation contours on the bedrock surface are shown in detail over areas of actual exposure or bedrock buried at shallow depth. Local well data published by Cushman (1964) were used to contour the bedrock surface in deeply buried areas. The contours show a glacially overdeepened, preglacial bedrock channel of the Connecticut River filled with buried outwash or ice-contact deposits. Figure 1b shows a complete Bouguer gravity map of the area. Noteworthy are the lack of gravity expression of the fault and the very clear expression of the buried bedrock channel as shown by the heavy dashed line representing the axis of a gravity trough. Bulk densities of the Triassic and Paleozoic bedrock are similar, explaining the lack of gravity expression of the fault. The major gravity lows on the map follow the axis of the known buried bedrock channel and where the latter is constricted, so too are the former. Quantitative determinations of thickness of the fill have not been made from the gravity data in this area, but detailed gravity studies in regions such as this one obviously are excellent for locating areas where unconsolidated deposits are unusually thick.

The apparent success of the gravity method in this area is chiefly because of the high density contrast between unconsolidated fill and bedrock. Bedrock densities vary between 2.55 and 2.92 gm/cc. In areas where the contrasts are low, the definition of narrow buried valleys similar to that in Figure 1 becomes more difficult. Some areas require application of data enhancement techniques, such as calculation of vertical derivatives or trend surface residuals of gravity, and some do not yield to the gravity method at all. A spectrum of examples can be seen by comparing the gravity maps in Figures 1b and 10b of this report with Figure 2 of Hall and Hajnal (1962), Plate 1 and Figures 8 and 9 of McGinnis, *et al.* (1963), Figure 6 of Heigold, *et al.* (1964), and Figure 15 of Lennox and Carlson (1967). Although these studies reflect varying degrees of success in applying gravity mapping to investigations of buried valleys, they also indicate the necessity of selecting the station spacing with care. One of them illustrates the unwarranted optimism that can be created by having enough borehole information to start essentially with a solution and adjust the gravity data until it clearly reflects this solution. In some areas gravity mapping of buried valleys works very well, in others it provides no help whatsoever to the hydrogeologist.

Gravity study of an intermontane basin. Gravity studies of intermontane basins, on scales much larger than those of buried valleys, have been made by many investigators. Of the 20 studies cited in a previous paragraph, none were made for hydrogeological purposes, but several provide data that could be so utilized. One of these (Von Huene, 1960) is presented here for discussion. It was followed by a detailed seismic refraction study (Zbur, 1963). The resulting data are shown in Figures 2, 3 and 12 of the present report.



EXPLANATION

Unconsolidated Quaternary sediments
Mostly Recent alluvium, but also includes local playa deposits, older alluvium, and lake sediments

Quaternary volcanic rocks
Predominantly basaltic

Tertiary sedimentary and volcanic rocks

Undivided Mesozoic and Paleozoic crystalline plutonic rocks

Contact

Fault
Dashed where approximately located; dotted where concealed

-10
Gravity contours
Hachures indicate area of low gravity. Interval 1.0 milligal

Seismic refraction line
Heavy bar indicates location of geophone spread; thinner line, row of offset shotpoints

0 20 40 60 80 MILES

N

Figure 2. Geology and residual Bouguer gravity map of Indian Wells Valley, California, after Von Huene (1960). Location of seismic refraction profiles from Zbur (1963). See Figure 3 for interpretations of profile A-A'.

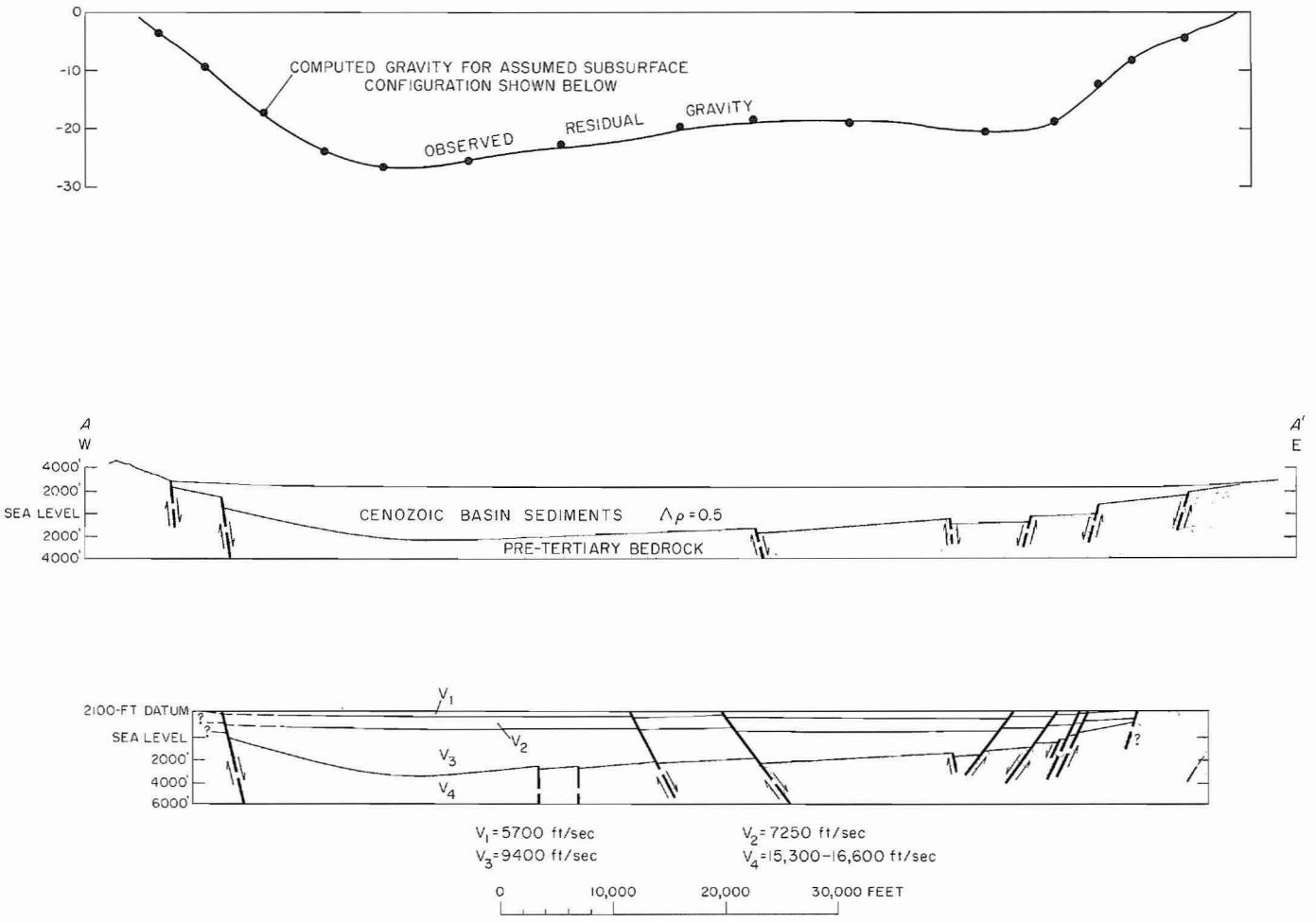


Figure 3. Residual Bouguer gravity profile in Indian Wells Valley, California, compared with gravity values computed from the model shown immediately below. At the bottom of the illustration is an interpretation of seismic refraction data along the same profile. After Zbur (1963, Fig. 19).

The map in Figure 2 shows Indian Wells Valley, California. This valley is bounded on the west, east, and southeast by ranges of relatively dense Mesozoic intrusive rocks and Paleozoic metamorphic rocks. Hills to the southwest are underlain by deformed Tertiary continental sedimentary rocks and others to the northwest, by Pleistocene lavas. The upper part of the valley fill includes Recent and older alluvium, Recent clay-rich playa sediments, and Pliocene to Pleistocene lake beds. At depth, older sediments (and possibly volcanic rocks) of Tertiary age underlie these Quaternary sediments. Approximately 1500 gravity stations (not shown) were occupied by Von Huene (1960) in an area of some 600 square miles, including the bordering mountain ranges.

Several features are apparent on the gravity map: (1) a large closed negative anomaly in the west-central area, a somewhat smaller one in the north-central part and a still smaller one in the east-central part, all of which overlie thick sections of valley fill; (2) gravity highs associated with all outcrops of pre-Tertiary basement rock; and (3) steep gravity gradients flanking the east and west sides of the valley. The gravity highs associated with outcrops of pre-Tertiary bedrock reflect, as do the gravity lows over the deeper parts of the valley, a pronounced contrast in density between the valley fill and bedrock. This, in turn, suggests relatively high porosities for the fill. The steep gradients flanking the east and west sides of the valley are expressions of high-angle faults separating Cenozoic from pre-Cenozoic rocks.

The locations of seismic refraction profiles established by Zbur (1963) are also shown in Figure 2. One, labeled A-A', has been selected for illustration in Figure 3, where it is shown with a coincident residual gravity profile and the interpretation made from both sets of data.

The interpretation of gravity profile A-A' suggests a maximum bedrock depth of approximately 1.43 kilometers (4700 feet), whereas the seismic data indicate a depth of 1.77 kilometers (5800 feet). The difference stems from the assumption of a homogeneous density contrast of 0.5 gm/cc for the valley fill. Actually, the density contrast between fill and bedrock must decrease downward and if allowances had been made for this decrease, the agreement in bedrock depth based on the two methods would have been better (Zbur, 1960, p. 76). The velocity stratification of the fill, where $V_1 < V_2 < V_3$, indicates a downward change in bulk density, which in turn suggests a change in porosity of the fill with depth. The combined use of both sets of data allows estimates of average total porosity. One

method for doing so is described in a later section of this paper and is illustrated in Figure 12.

Unlike buried channel deposits of glacial and glaciofluvial origin, the Cenozoic strata in Indian Wells Valley and other intermontane basins are generally sheet-like in configuration and most extend across the width of the valley, varying only slightly in thickness. It is not suggested that intermontane deposits do not display abrupt lateral variations in density or velocity, for in some areas they do. By and large, however, deposits in such basins display greater stratigraphic and lateral continuity than do mixed glacial and fluvial sediments in buried valleys because the conditions of origin were less variable in space and time. For this reason they are somewhat more amenable to quantitative gravity interpretation than buried valley deposits.

Seismic case histories

Seismic studies of a buried river channel. Figure 4 illustrates an area in New Jersey where a combination of borehole and seismic refraction data were used by Bonini and Hickok (1958) to map a buried bedrock channel carved in Triassic sedimentary rocks. The objective of the study was the preparation of a detailed bedrock contour map to aid in the selection of water well sites. The unconsolidated sediments overlying the Triassic bedrock include glacial outwash sand and gravel, lacustrine silt and Recent alluvial sediments. Compressional wave velocity contrasts in the area are high, the unconsolidated sediments displaying a range from 0.27 km/sec (900 ft/sec) to 2.07 km/sec (6800 ft/sec) and the bedrock, 3.23 km/sec (10,600 ft/sec) to 5.00 km/sec (16,400 ft/sec). Standard recording and computational techniques were employed and the results, as confined by borehole data, were highly successful. According to the authors (Bonini and Hickok, p. 488), the average error in seismic prediction of bedrock depth in their study was only 7 percent.

A somewhat different objective is illustrated by the study of Johnson (1954) in a glaciated area of Illinois to map stratigraphic horizons *within* the Pleistocene sediments. The method, as checked by borehole data, was particularly successful in distinguishing glacial drift of Wisconsin age from that of Illinoian age. Travel-time curves displaying delayed arrivals were compiled in areas where water-bearing strata of sand and gravel with relatively low velocity were present within the stratigraphic section, thus providing a potential guide to more porous unconsolidated material.

Seismic studies of intermontane basins. Most seismic investigations of intermontane basins differ in scale, field method and interpretation procedure, from those of buried valleys. Long geophone spreads are employed, and in-line shots are frequently separated from the ends of the spreads by thousands to several tens of thousands of feet. Because of the simpler layering and, on this scale, the more nearly planar nature of the refractors (see, for example, Figure 3), conventional intercept-time methods generally are used for interpretation instead of the graphical or time-term methods (Hawkins, 1961; Pakiser and Black, 1957; Willmore and Bancroft, 1960) applied in buried valley studies. The results of such work generally accord well with data from borings where the latter are available for comparison.

Most seismic investigations of intermontane basins in the western United States reported in the literature have revealed

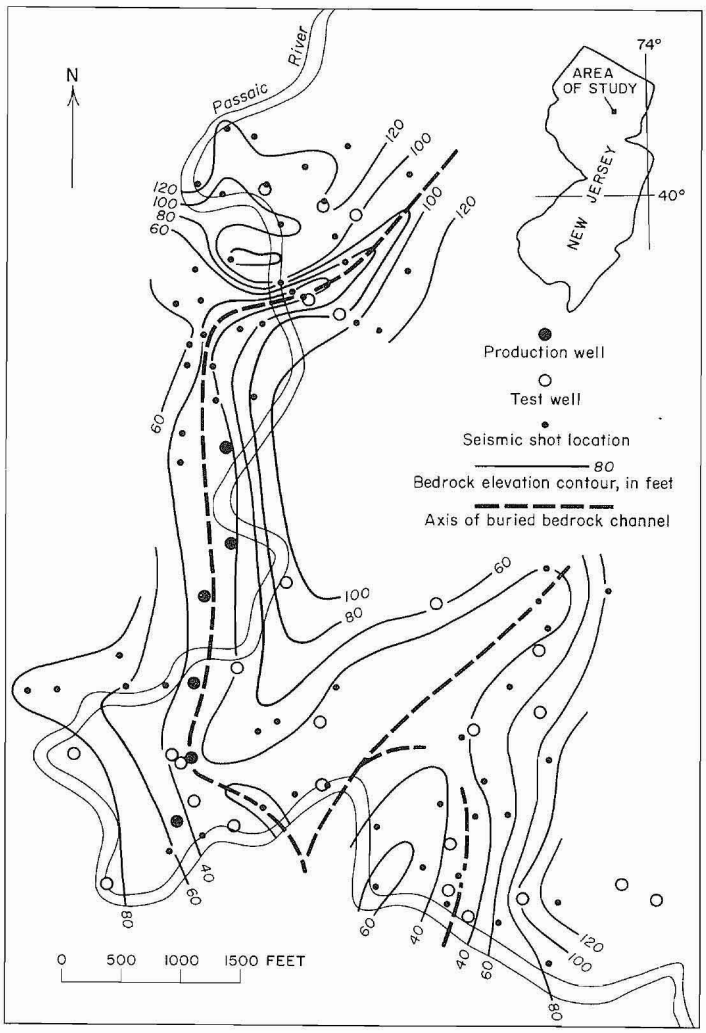


Figure 4. Buried bedrock topography beneath the Passaic River, north-central New Jersey, as determined from combined seismic refraction and borehole measurements by Bonini and Hickok (1958). These data were used to site production wells for a city water supply.

velocity stratifications similar to that shown in Figure 3; hence that profile serves to illustrate the geologic conditions generally encountered (see, also, the data of Table IB). Here, as in most studies of buried valleys, the principal objectives in the past have been determination of the configuration of the bedrock surface and analysis of the velocity stratification of the basin fill. Although the seismic method is well suited for achieving these objectives, potentially it provides additional information on porosity that could be used by the hydrogeologist. The intermontane basin study by Zbur (1963) is employed in a later paragraph to illustrate this point.

Seismic determination of groundwater level. It has long been observed, in the course of normal refraction studies of medium- to coarse-grained unconsolidated deposits, that compressional wave velocities increase sharply at the water-table. The zone of saturation is an excellent refractor the upper boundary of which can be determined with considerable accuracy by the seismic

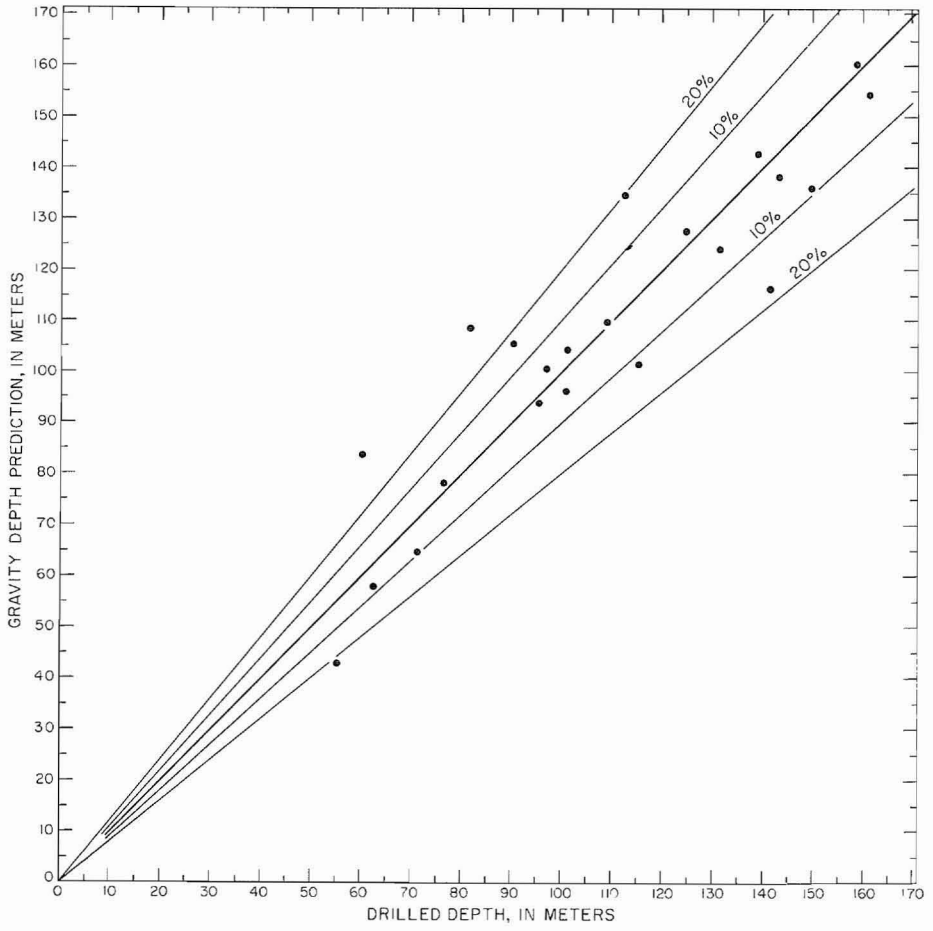


Figure 5. Comparison of 22 gravity determinations of depth to bedrock beneath unconsolidated alluvium, with borehole determinations of depth at the same locations. Area is in the Santa Ana River Valley, north of Riverside, California. Heavy line reflects perfect accuracy in prediction, lighter lines, errors of 10 and 20 percent. Data from Eaton, *et al.* (1964).

method. An excellent review of this and related phenomena, plus a discussion of several criteria whereby one can distinguish the water-table from other seismic interfaces of similar velocity contrast, has been presented by Levshin (1961). Although further research is clearly desirable, application of the techniques he suggests–for example, combined use of compressional and shear wave measurements and study of frequency distribution and attenuation characteristics–should make it possible to map the water-table in heterogeneous unconsolidated deposits of varied lithology in many regions, particularly when used with surface electrical measurements.

Accuracy and error in shallow gravity and seismic depth determinations

Accuracy of depth determinations by gravity. Few of the studies cited above involved efforts to measure, from gravity data, the thickness of unconsolidated fill in channels and buried valleys. In most studies, the primary aim was to trace the course of buried valleys or to delineate subsurface groundwater barriers in the form of buried bedrock ridges. The gravity method does not lend itself nearly as well to quantitative depth determinations on this scale as does the seismic refraction method and there are few reports that evaluate its accuracy in the unconsolidated environment. Eaton, *et al.* (1964, p.12-15) attempted such an evaluation, the results of which are shown in Figure 5. For comparison, seismic data of a similar nature are shown in Figure 6.

The gravity measurements of Figure 5 were made in the valley of the Santa Ana River, north of Riverside, California. Bedrock consists of relatively dense plutonic rock and is overlain by poorly consolidated alluvium. The configuration of the bedrock surface had been defined by more than 40 boreholes sunk during an exploration program for cement limestones. Twenty-two of these holes, with bedrock depth ranging from 55.2 to 161.3 meters (179 to 523 feet), were chosen as sites for gravity stations. Additional stations were located on bedrock in the general vicinity to determine the nature of the gravitational field arising from variations within bedrock. Bouguer values at the latter stations were contoured and the derived gravity field subtracted from that measured at the borehole stations. Residual anomaly values at the borehole stations were presumed due only to variations in the thickness of the alluvium. Actually, an additional source of variation exists in this area, for large volumes of rock have been removed during underground mining for limestone, several hundred feet below the surface. The cavities so created constitute local deficiencies of mass for which corrections should have been made, but were not.

A plot of residual Bouguer gravity values versus depth to bedrock was fitted by the method of least squares. In the preparation of Figure 5, the fitted curve was used to predict bedrock depths from values of residual gravity and these are compared with actual borehole depths. There is considerable scatter, but the *average* error in correlation is approximately 10

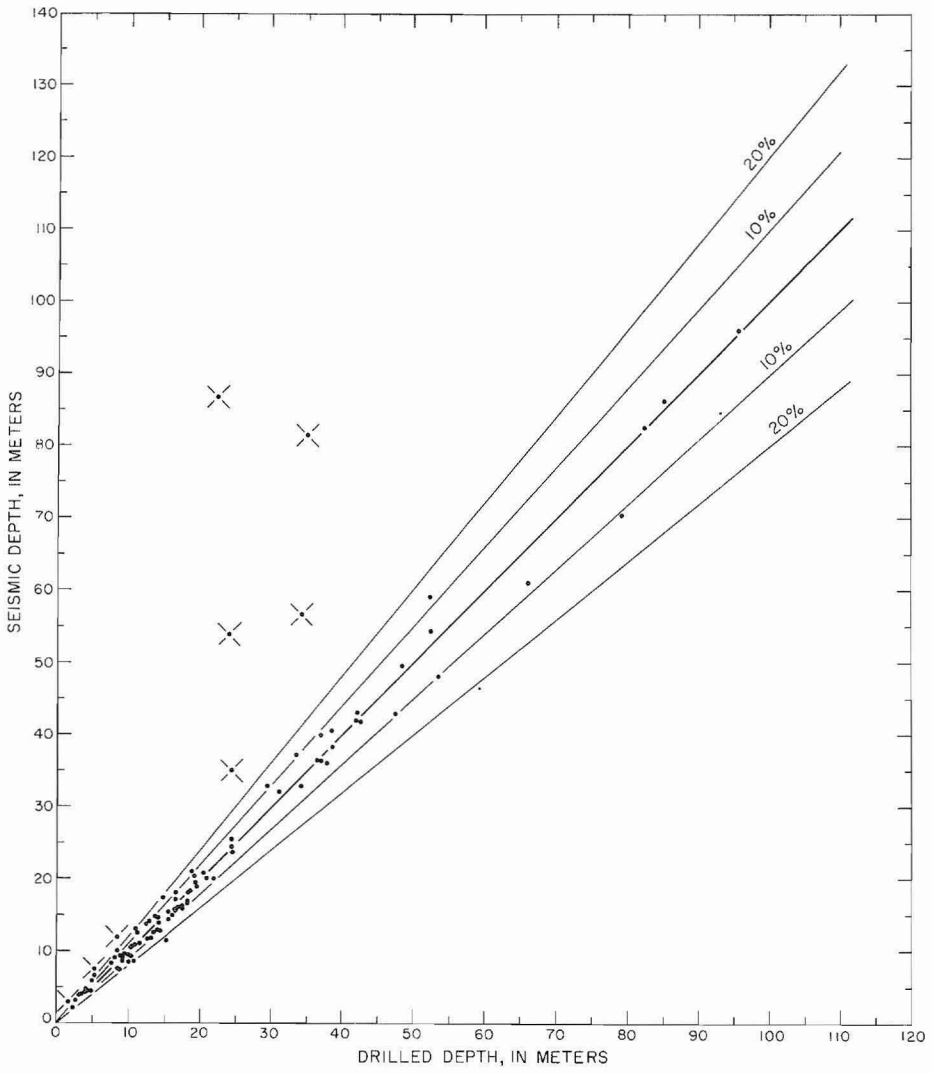


Figure 6. Comparison of 97 seismic refraction determinations of depth to bedrock beneath unconsolidated alluvium and glaciofluvial sediments, with borehole determinations of depth at the same locations. Heavy line reflects perfect accuracy in prediction; lighter lines, errors of 10 and 20 percent. Starred data points represent errors greater than 30 percent. Data from Bonini and Hickok (1958), Brown and Robertshaw (1957), Hasselström (1951), Lennox and Carlson (1967), McGinnis and Kempton (1961) and Welin (1958).

percent. The shallowest well utilized in this study has a bedrock depth of 55.2 meters (179 feet). Shallower wells were not used because of the known rapid increase in density in the upper 45 meters (150 feet) or so of alluvial deposits (Eckis and Gross, 1934, p. 105 - 108).

Accuracy of depth determination by seismic means. Figure 6 is a graphic plot of bedrock depths determined by drilling versus depths predicted from seismic refraction measurements. The data represent measurements made in glaciated areas of Canada, England, Sweden and the United States. Most of the depth estimates fall within the 10 percent limits of actual depth and the average scatter is less than that for the gravity depth predictions shown in Figure 5. Nevertheless, the magnitude of error for certain of the seismic determinations is large, indicating that some areas are less amenable to seismic study than others. Sources of such errors are outlined as follows:

Sources of error in depth determinations. Several assumptions are made in the interpretation of seismic refraction data and from these arise some of the errors encountered in interpretation. For example, we assume that Fermat's principle holds, that $dv/dz \geqslant 0$, and that within a given stratigraphic unit, $dv/dx = dv/dy = 0$. In practice, further constraints must be imposed by the following idealizations: (1) the body consists of a small number of more or less horizontal elements separated by plane or gently curving velocity discontinuities, and (2) $dv/dz = 0$, except at interfaces where abrupt velocity discontinuities occur. In most field investigations, Fermat's principle holds as long as the layers are thick compared to the wavelength of the seismic waves. Thin layers distort the waves when nearly equal to the wavelength, but where very thin, they have little effect.

In certain settings, $dv/dz \leqslant 0$. Two examples, commonly encountered in groundwater investigations, are high-velocity glacial till overlying lower velocity sand, and the presence of caliche within the zone of unsaturated alluvium in arid and semiarid intermontane basins. Other examples of geological situations which violate the basic assumptions listed above are tabulated in Table VA for seismic and Table VB for gravity studies.

Applications of gravity and seismic methods in estimating porosity of unconsolidated sediments

Sediment porosities can be measured *in situ* by borehole logging techniques. The emphasis in this paper, however, is on surface

Contents

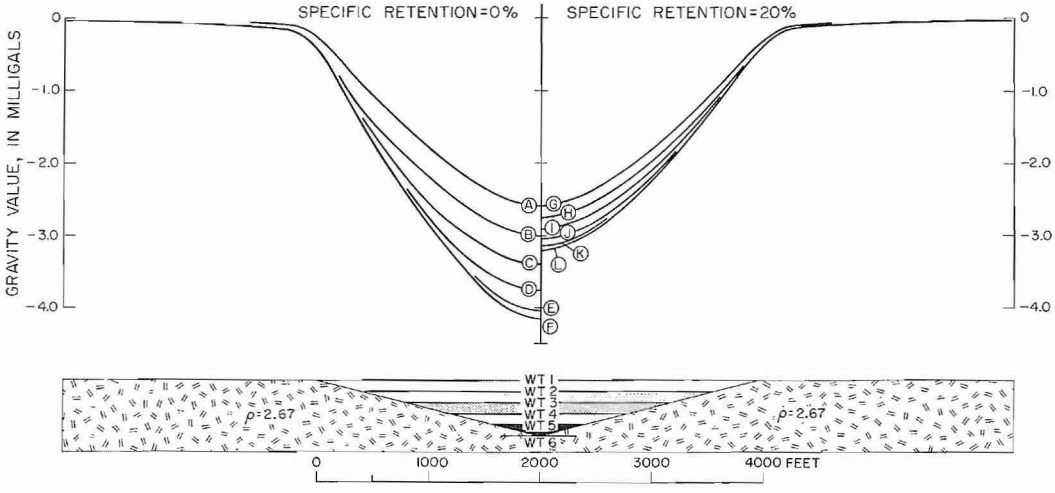


Figure 7. Cross section of idealized bedrock channel carved in granitic rocks and filled with unconsolidated sediments, plus the Bouguer gravity profiles associated with it. Bedrock density = 2.67 gm/cc; sediment grain density = 2.67 gm/cc and porosity = 0.33. Shown above the cross section are gravity anomaly curves corresponding to various positions of the water table (*WT*). Curve *A* is associated with *WT*1, *B* with *WT*2, etc. The gravity data reflect a specific retention for the sediments of 0.0 on the left and 0.20 on the right.

geophysical measurements. They are regarded not as substitutes for the logging methods, but rather as methods to be applied in the early stages of aquifer evaluation. Water wells may be present in the area of interest at this stage, but they may be too shallow for meaningful pumping tests or logging, or it may be that the cost of such tests is prohibitive. Under these circumstances, surface geophysical methods can be used to aid in aquifer evaluation. However, high porosities, if found, are of value only where permeability is also high. Neither of the methods described as follows makes possible an estimate of permeability at this time.

Determination of average total porosity by the gravity method

Gravity variations in an idealized model. Surface measurements of gravity can be used to estimate, at least approximately, the average total porosity of alluvial fill *in situ*, provided that the fill is fairly homogeneous, the lower surface of the fill rests on material of appreciably higher density and broad relief, and the geometry of the fill is independently known from drill-hole or accurate seismic refraction data. Measurements of this sort, where made in detail, can aid in the evaluation of intermontane basin or buried valley deposits as storage facilities for water. Similar measurements made during periods of seasonal or long-period water-level decline, may also provide a means of estimating, at least to a first approximation, the specific yield of the fill. Both applications are described.

To estimate porosity, the geometry of the fill and its total gravitational attraction are utilized to determine the density contrast between fill and underlying material. In essence, the fill is 'weighed' in place, *en masse*, and when its average bulk density has been determined, its porosity is calculated from additional data bearing on its grain density. A primitive version of the technique has been described by Eaton, *et al.* (1964, p. 15 - 21).

An example of the method, is given in Figure 7 which shows Bouguer gravity profiles for an idealized channel-fill aquifer of triangular cross section. This channel is assumed to have infinite extent in a direction normal to the plane of the page and variable, but high (.55 to .88 gm/cc) density contrasts between the fill and surrounding bedrock. Such contrasts are well within the limits suggested by the data of Tables II and IV, particularly for the case of a pre-Tertiary crystalline bedrock. No regional gravity gradient is present.

Two families of gravity profiles are shown. Profiles *A* through *F*, on the left, represent anomalies associated with an idealized horizontal water-table of varying elevation in a material with negligible specific retention. As the upper surface of the zone of saturation falls from level 1 to level 6, the water in the interstices of the alluvium drains downward and none is retained in the sediments above the water-table to contribute to their mass. When the water-table is coincident with the surface of the ground, the maximum value for the negative gravity anomaly associated with this channel is slightly less than 2.6 milligals. With the aquifer completely drained, the maximum value of the anomaly is approximately 4.1 milligals, a difference of about 1.5 milligals. For water levels between the top and base of the aquifer, the value of the associated negative anomaly lies between these two limiting values, as shown in the illustration.

Table V. Sources of error in seismic and gravity studies of deposits of unconsolidated sediments.

(A) *Seismic refraction method*
- (1) Insufficient velocity contrast at hydrologically important interfaces.
- (2) Lateral, discontinuous velocity variations in either the unconsolidated material or the underlying bedrock, including the existence of discontinuous lenses and beds. Such variations are particularly troublesome in glacial and glaciofluvial sediments.
- (3) Velocity anisotropy within the unconsolidated sediments.
- (4) Velocity inversions, the corrections for which require borehole velocity information and knowledge of local stratigraphic thicknesses.
- (5) The occurrence of velocity layers too thin to be observed in a separate branch of the travel time curve.
- (6) A highly fractured or weathered bedrock surface with anomalously low velocity.
- (7) Irregular, curved, or steeply dipping refractors making accurate determinations of true velocity difficult.
- (8) Errors in the values assumed for velocity and thickness of the 'weathered' layer, if not measured, or unrecognized variations in these parameters, if measured at only a few scattered points.
- (9) Anomalous or highly disturbed conditions at the shot hole.

(B) *Gravity method*
- (1) Errors in determination of elevation of stations.
- (2) Incorrect calibration of the meter or unrecognized lack of linearity in its response.
- (3) Incorrect densities used in Bouguer and terrain corrections.
- (4) Improper evaluation of the regional gradient.
- (5) Pronounced irregularity of the bedrock surface.
- (6) Lateral density variations in either the unconsolidated fill or bedrock.
- (7) Variations of density with depth.

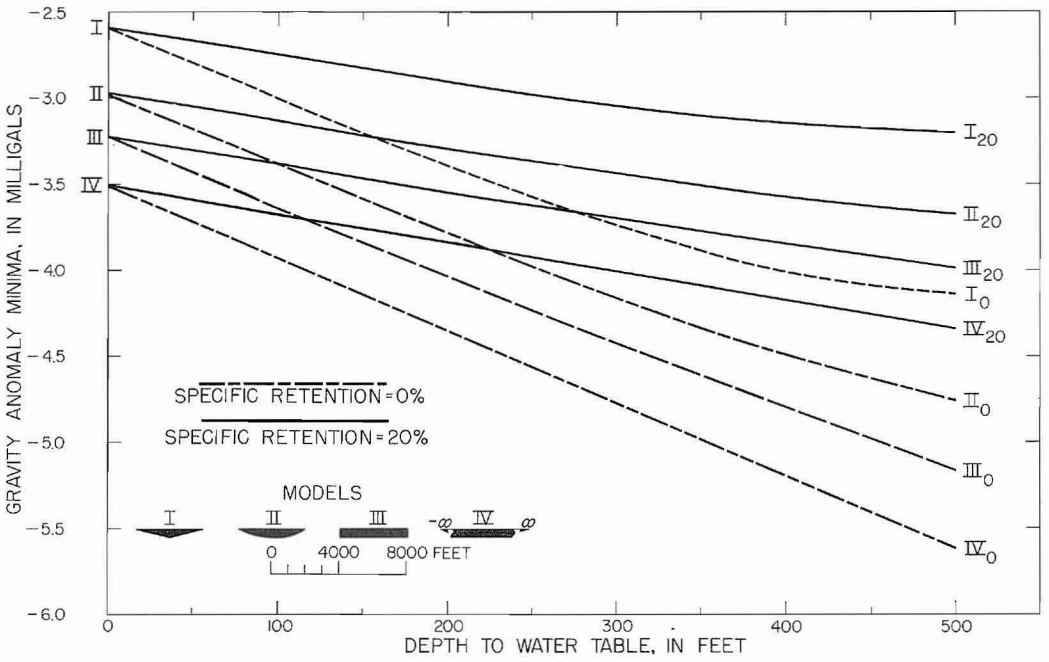


Figure 8. Maximum gravity effects of declining water-tables in four aquifers with different cross-sectional configurations. Two curves are shown for each aquifer, one for specific retention (*SR*) of 0.0, the other for 0.20.

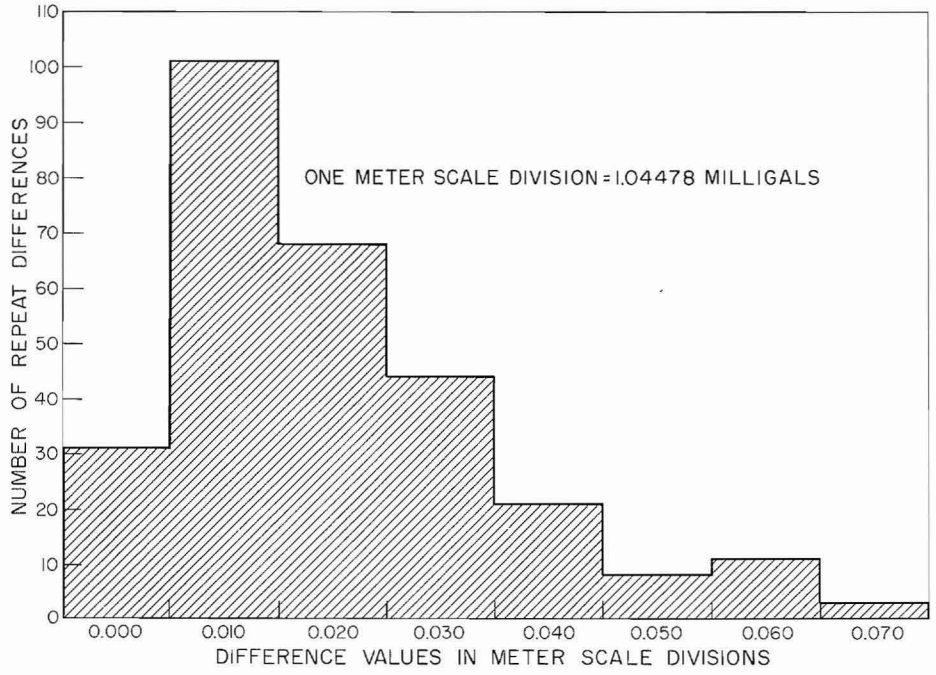


Figure 9. Histogram showing precision of gravity measurements with a LaCoste and Romberg Geodetic Model gravity meter for 287 repeat observations on alluvium and bedrock on the South Fork of Owl Creek, Hot Springs County, Wyoming. Tidal and drift corrections made before plotting.

Profiles *G* through *L* on the right, represent gravity anomalies associated with varying water levels in an alluvium-filled channel of identical configuration, but with a fill having a specific retention of 20 percent. Here, the maximum total gravity variation associated with a drop in the water-table from the surface to the base of the aquifer is 0.61 milligal, less than half the variation for the aquifer with negligible specific retention. The effect of specific retention is thus seen to be significant. In both instances, the anomaly associated with this aquifer could be detected readily in the course of a gravity survey with closely spaced stations. In addition, a distinction could be made between shallow and deep levels of the water-table if measurements were made over the aquifer both at the beginning and at the end of a period of pronounced water-level change.

Figure 8 illustrates gravitational effects resulting from a decline in water level in four aquifers of different cross-sectional configuration. This illustration shows the maximum gravity effect of a declining water-table in aquifers of like porosity and thickness, three with similar width but different cross-sectional configuration, and one with infinite lateral extent. Two curves are presented for each aquifer, one representing a specific retention of 0 percent, the other, 20 percent. Variations arising from differences in shape for a given specific retention and from differences in specific retention for a given cross-sectional shape are both notable.

A water-table decline of 100 feet in an infinitely extending aquifer with a porosity of 0.33 and negligible specific retention produces a change in the observed Bouguer gravity field of 0.42 milligal. A change of approximately 0.17 milligal results from a similar decline in an infinitely extending aquifer with a specific retention of 20 percent. If precise gravity measurements were made periodically in an area where the annual fluctuation of the water-table was around 100 feet, or where, over a period of years, water levels declined by this amount as a result of inadequate recharge, they would provide a means of estimating the specific yield of the materials drained.

The preceding discussion and Figures 7 and 8 serve to indicate the general magnitude of gravitational variations to be expected from changes in groundwater levels observed in alluvial aquifers. These variations are easily detectable with a modern gravity meter, as shown by Figure 9. The data for this figure were obtained by one of the writers for determining the reproducibility of a gravity meter being used to map very small gravity anomalies. Figure 9 shows the frequency distribution of 287 repeat gravity observations made in the valley of the South Fork of Owl Creek, Wyoming, during the late fall of 1964. Ambient temperatures fluctuated widely. Two operators made the measurements at different times with the same LaCoste and Romberg Geodetic Model gravity meter. The modal value of the repeat differences in this study is approximately 0.01 milligal and the great bulk of the differences are as small as, or smaller than, 0.03 milligal. From the standpoint of detectability and precision, the measurement of a gravity variation of 0.17 milligal for hydrological purposes does not constitute a serious problem. However, the *accuracy* of the measurements depends on the accuracy of calibration of the instrument and this requires close and continued attention.

Methods of interpretation. The basis for estimating the porosity of an alluvial aquifer from gravity measurements involves measuring three quantities: the residual gravity anomaly associated with the aquifer, the density of the material on which the aquifer rests, and the average grain density of the aquifer. The residual gravity anomaly is used with the known geometric configuration of the aquifer and its water-table, determined from well data or by seismic refraction surveys, to calculate two density contrasts between the aquifer and the material on which it rests.* One density contrast is for the saturated part of the aquifer and the other, the unsaturated part. The algebraic differences between each of these contrasts and the density of the bedrock, which must be measured independently, yield values for the unsaturated and saturated bulk densities of the alluvium, respectively.

The relationship between dry bulk density ρ_d, porosity ϕ, and average grain density ρ_g is

$$\phi = \frac{\rho_g - \rho_d}{\rho_g} = 1 - \left(\rho_d / \rho_g\right) \qquad 1$$

This equation is the basis for conventional methods of measuring rock porosity in the laboratory (Manger, 1963) and it is also the basis for the gravimetric method proposed here.

*The presence of wells in an area does not mean necessarily that evaluation of an aquifer is complete. In many areas, wells drilled for the purpose of irrigation or urban water supply, and penetrating the bedrock surface, provide data on bedrock depths, but an estimate of total average porosity may still be needed for further development of the region. Such was the case for the illustrative example shown in Figure 10.

For an aggregate saturated with a fluid of density ρ_f, its saturated bulk density ρ_s is related to its dry bulk density and porosity as follows:

$$\rho_s = \rho_d + \rho_f \cdot \phi \qquad 2$$

In practice, either of two standard gravity interpretation methods can be used to determine the density contrast between the alluvial fill and the underlying material. If the fill occupies a linear channel or elongate trough that can be regarded as a two-dimensional body, a graticule can be employed to calculate a theoretical gravity profile from the known cross-sectional configuration of the body. If the body is not two-dimensional, more sophisticated integration techniques are required employing a high-speed digital computer.

Alternatively, Gauss's Theorem provides a means of determining the amount of disturbing mass responsible for a given gravity anomaly. The volume of the body can be calculated from its known geometry and the bulk density calculated, in turn, from the ratio of mass to volume. LaFehr (1965) investigated the effects of both depth and cross-sectional configuration of a gravitating body on the calculation of its total mass by Gauss's Theorem. He showed that the accuracy of the method is high, especially if the ratio of width to thickness of the body is ten or greater. Cursory examination of published data on both drift-filled buried valleys and intermontane valley-fill deposits indicates that this ratio of width to thickness is commonly met or exceeded (Mabey, 1960, Figs. 25, 28 and 31; Hall and Hajnal, 1962, Figs. 4 and 5; and McGinnis, *et al.*, 1963, Fig. 5).

The operationally simpler of the methods is the graticule analysis. Because one of the illustrative examples presented at the end of this section of the paper involves this method, its adaptation is outlined briefly in the paragraphs that follow.

Hubbert (1948, p.217 - 219) showed that the gravitational attraction dg at the origin of an x-y-z coordinate system, because of an infinitely long prism parallel to the y-axis, bounded by the planes z and $z + dz$, and lying between two inclined planes which intersect at the y-axis and make with the plane z=0 dihedral angles of θ and $\theta + d\theta$, respectively, is

$$dg = 2k\rho\, dz d\theta \qquad 3$$

where k is the constant of gravitation and ρ, the volume density of the prism. This equation is the basis for the computation of the gravimetric effect of a two-dimensional body of arbitrary cross section, the body being divided into trapezoidal cells by radial lines with constant separation $\Delta\theta$ and by horizontal lines with constant separation ΔZ. If the volume density is uniform throughout the body, the equation for the calculation of the total gravitational attraction of the body can be reduced (Hubbert, 1948, p. 219 - 223) to

$$g = 2k\rho\ \Delta Z\ \Sigma\theta \qquad 4$$

To calculate the gravitational attraction of an elongate body of alluvium, the lower part of which is saturated with water, and the upper part, dry, or nearly dry, the body is divided horizontally into two parts which display density contrasts with

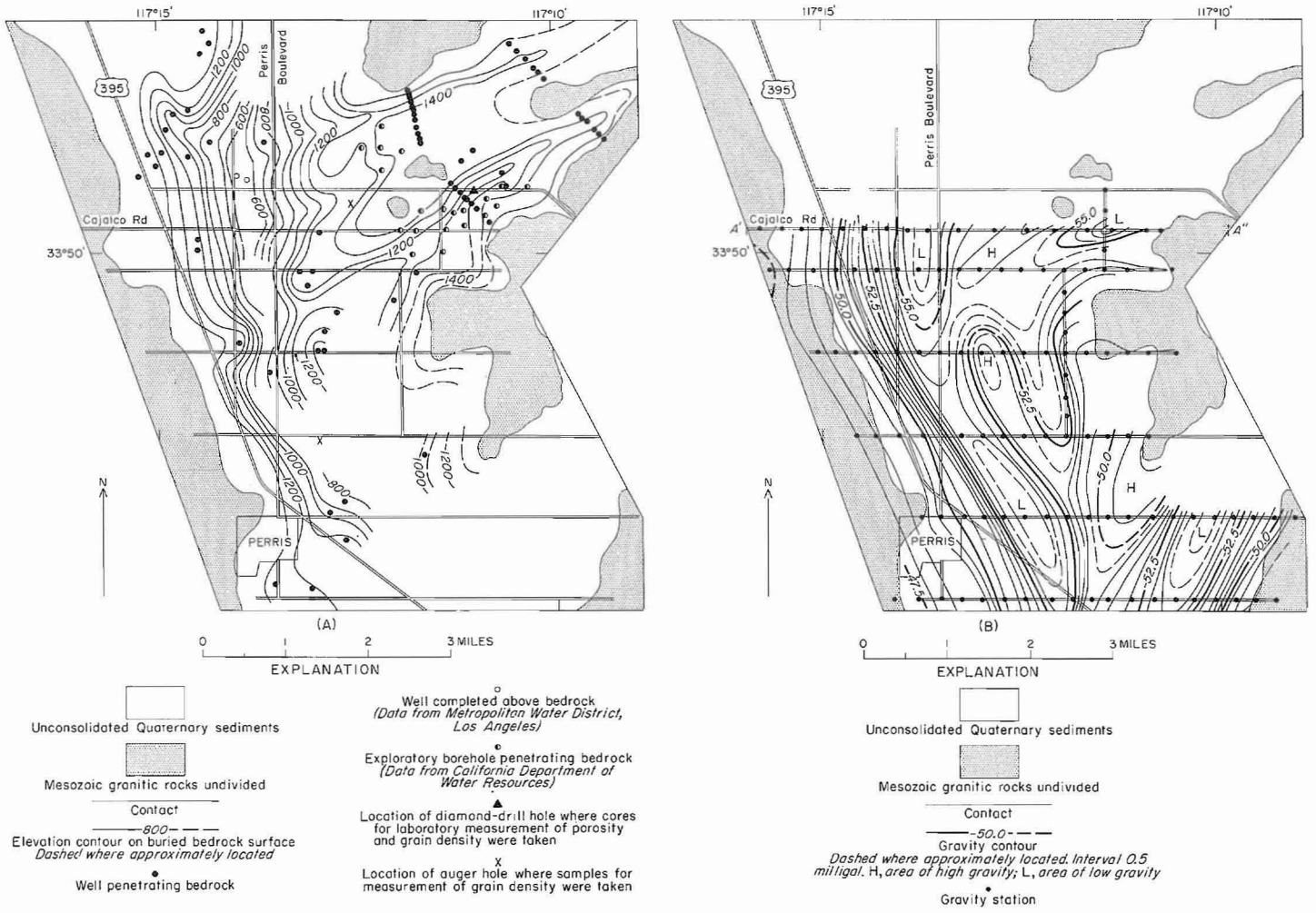


Figure 10. (a) Generalized geologic map of a portion of Perris Valley, California, showing distribution of bedrock outcrops and structure contours on the buried bedrock surface. (b) Complete Bouguer gravity reconnaissance map of the same area.

the underlying bedrock of $\Delta\rho_s$ and $\Delta\rho_d$, respectively.* Under these conditions Equation 4 can be written

$$g = 2k\Delta Z\left[\Delta\rho_s\ \Sigma\theta_s + \Delta\rho_d\ \Sigma\theta_d\right] \qquad 5$$

where $\Sigma\theta_s$ and $\Sigma\theta_d$ refer to independent summations over the saturated and unsaturated parts of the body, respectively.

Because graphical integrations of this sort break down when the origin of the graticule is placed at the edge of the area being measured, the uppermost part of an aquifer has to be treated independently. The simplest means of doing this is to approximate it with prismatic slabs of analytical shape, calculating the numerical value of gravitational attraction for these slabs, and adding their contribution to the attraction calculated graphically for the rest of the body.

*It is not necessary to assume that the part of the aquifer above the water-table is completely dry, but if it has appreciable specific retention, the value for ρ_d, which is then the density of partially wetted alluvium, will cause an error in the final result. The magnitude of this error depends on the value of the specific retention and the relative thickness of that part of the aquifer above the water-table.

If the density of the bedrock on which the alluvium rests is designated ρ_r, the density contrast $\Delta\rho_d$ between bedrock and dry alluvium can be written, upon substitution from Equation 1

$$\Delta\rho_d = \rho_r - \rho_d = \rho_r - \rho_g + \rho_g \cdot \phi \qquad 6$$

Similarly, the density contrast $\Delta\rho_s$ between bedrock and saturated alluvium can be written, upon substitution from Equations 2 and 1

$$\Delta\rho_s = \rho_r - \rho_s = \rho_r - \rho_d - \rho_f \cdot \phi \qquad 7$$

Substituting Equations 6 and 7 in 5 and solving for ϕ, we get

$$\phi = \left[\frac{(g/2k\Delta Z) - (\rho_r - \rho_g)\ (\Sigma\theta_s + \Sigma\theta_d)}{\rho_g\,(\Sigma\theta_s + \Sigma\theta_d) - \rho_f\,\Sigma\theta_s}\right] \qquad 8$$

which is an algebraic expression for the porosity of a two-dimensional body of alluvium under water-table conditions, expressed in terms of its associated gravity anomaly g, its grain density ρ_g, the density of the bedrock on which it rests ρ_r, and the cross-sectional configuration of both its saturated and unsaturated parts ΔZ, $\Sigma\theta_s$, and $\Sigma\theta_d$.

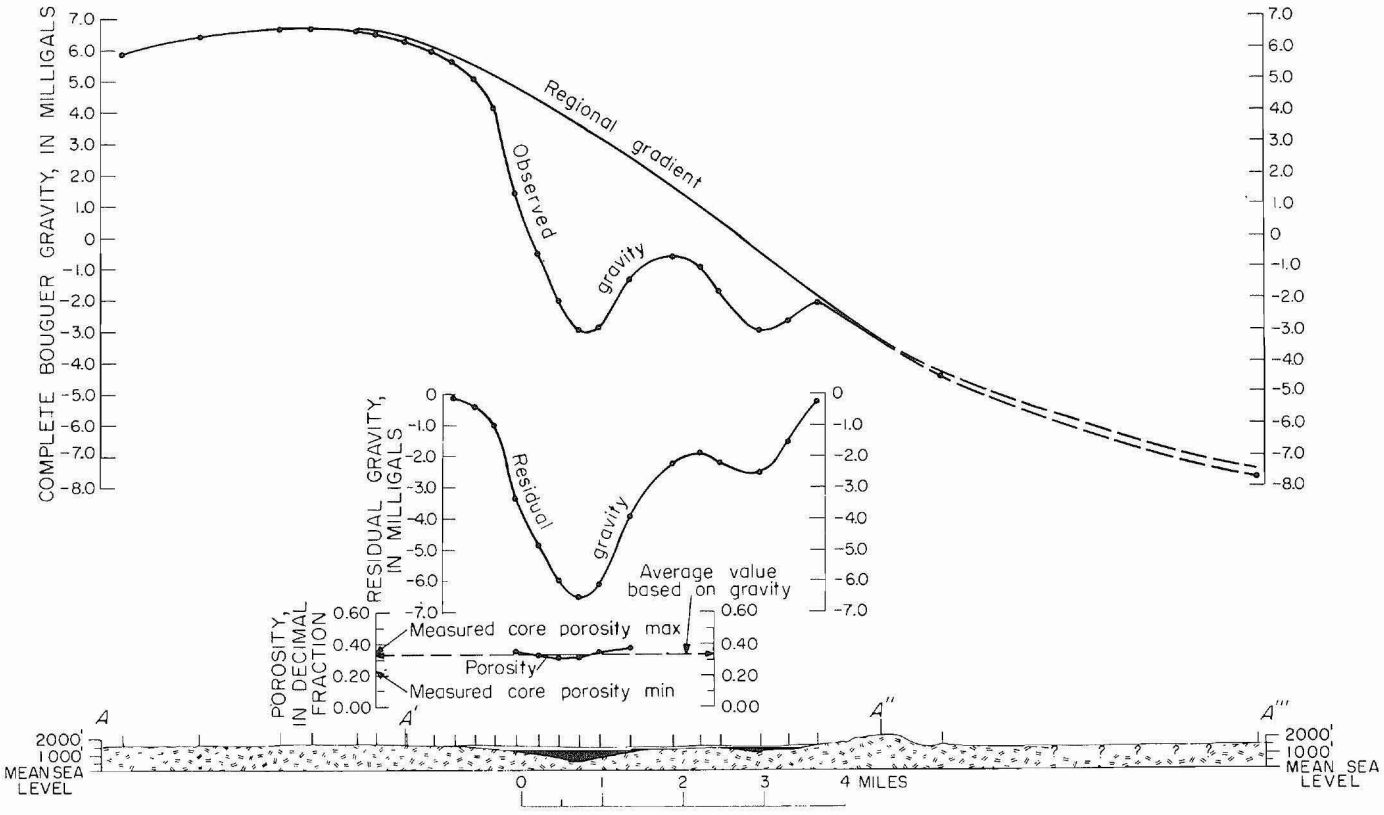


Figure 11. Profiles of observed Bouguer gravity, residual gravity, and porosity, across Perris Valley, California. That part of the cross section labeled A'-A'' corresponds to the Cajalco Road traverse of Figure 10.

Sources of error. Possible sources of error in estimating porosity by the suggested gravity method include: (1) spatial variations in porosity and density of the alluvium or underlying bedrock, (2) gross inaccuracy in the knowledge of the configuration of the bedrock surface or the water-table, (3) inability to distinguish that part of the gravity field associated with the alluvial fill from other sources contributing to the total field, and (4) incorrect selection of densities for use in data reduction.

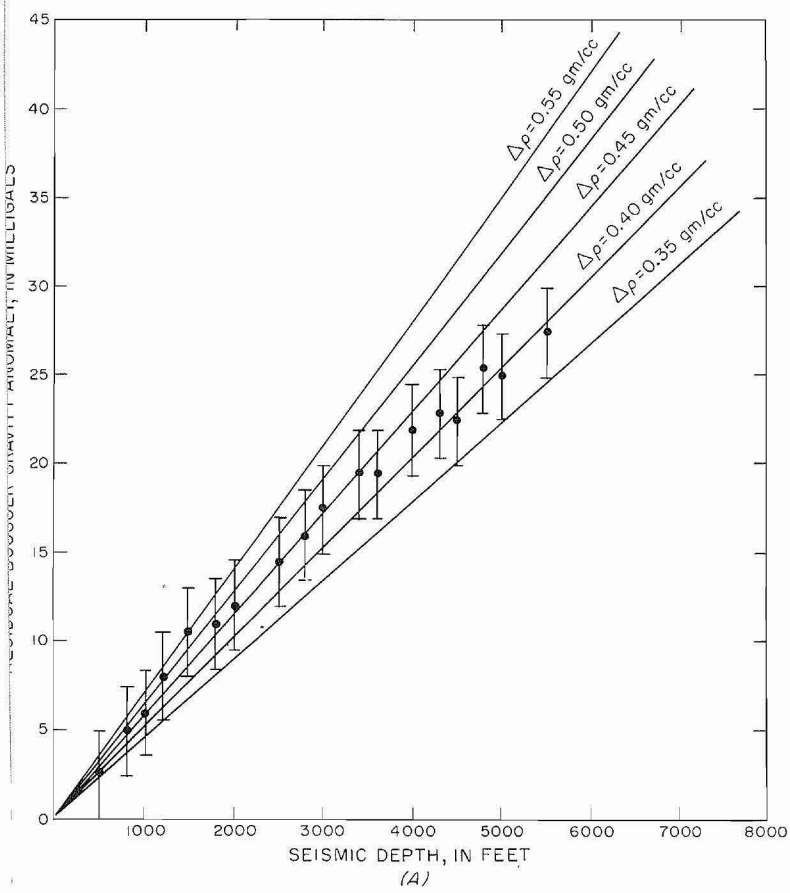
Of the several sources of error just mentioned, that arising from lithofacies variation is the most difficult to evaluate and probably poses the greatest potential difficulty in applying the method proposed. In general, the application of these methods will be more satisfactory in the case of valleys filled with homogeneous alluvium than for those filled with glacial and glaciofluvial sediments. However, values of porosity determined in this way are at best only estimates. They represent averaged values for the deposit as a whole. Nevertheless, they may be useful in some studies of glacial or preglacial valleys even under these conditions.

An example for the buried valley model. An application of the method just described is illustrated in Figures 10 and 11. It furnishes some idea of the magnitude of error involved by comparing porosity values determined from gravity measurements with values determined from drill cores by conventional laboratory techniques. The area was chosen for study both because the geometry of the alluvial aquifer was well known from drill-hole data and because rotary drill samples of both the aquifer material and the underlying granitic bedrock were available. The area is part of Perris Valley, southeast of Riverside, California.

Figure 10a shows the distribution of outcrops of bedrock in the area, as well as elevation contours on the buried part of the bedrock surface. The elevation contours, taken from an unpublished map of the Metropolitan Water District of Southern California, have been modified on the basis of 20 new bedrock borings, the data for which were provided by the California Department of Water Resources. Figure 10b is a reconnaissance Bouguer gravity map of part of the same area which was contoured without reference to the bedrock map. The correlation between the two sets of data is excellent.

Traverse A'-A'', along Cajalco Road, was chosen for a test of the porosity estimation method just described. It was selected because it crosses the axis of the principal buried valley perpendicularly, because well data are more numerous in the general vicinity of this traverse than elsewhere in the area, and because diamond-drill cores of the fill were available from a test boring immediately to the north.

The gravity profile used to estimate porosity was extended beyond the borders of the map area a distance of 6.5 to 8.0 kilometers (4 to 5 miles) in each direction, to evaluate the regional gradient. The *extended* profile, A-A''', is shown in Figure 11 including the section A'-A'' that falls within the borders of the map area (Figure 10b). New station locations within A'-A'' were occupied when the longer profile was surveyed, resulting in a different elevation datum and different densities being used in the reduction of the data. Thus Bouguer gravity values on the map (Figure 10b) and profile (Figure 11) are not the same. For the



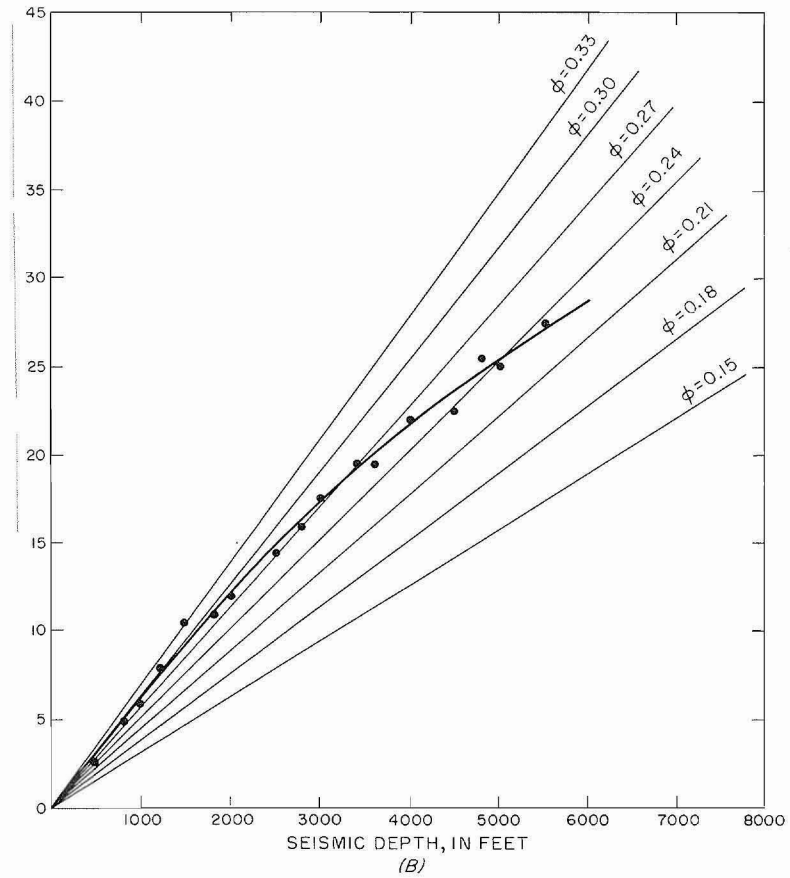


Figure 12. Plots of seismic depth determinations in Indian Wells Valley, California, versus residual Bouguer gravity values at the same locations. In (a), the data are shown against a field of curves indicating the gravitational attraction of infinite, horizontal slabs of density contrast as indicated; in (b) the data are shown against a field of curves indicating the gravitational attraction of similar slabs of porosity as indicated.

profile A-A''', variable densities were used for Bouguer corrections, while for the reconnaissance map, a single density was used.

Below the observed gravity data in Figure 11 is a geological cross section prepared from Figure 10a extended east and west of the map area by geological reconnaissance. Data on the water-table have been added. The geometry of this section was used with residual gravity values at six stations over the central part of the valley to calculate porosities for the alluvial fill using Equation 8. Additional data necessary for the solution of the equation included measurements of grain density (58 measurements, average value: 2.675 gm/cc) collected at three localities shown in Figure 10a and bedrock density (30 measurements, average value: 2.735 gm/cc) determined from drill cores collected on both sides of the valley. The density of the ground water in the area was assumed to be 1.00 gm/cc.

The results of these calculations are shown as heavy dots on a porosity profile in the lower part of Figure 11. The values range from 0.308 to 0.372, with an average of 0.334. As a check on the method, 10 semiconsolidated samples of the fill gathered at depths ranging from 6 to 82 meters (20 to 270 feet) from a boring one half mile north of the east end of A'-A'' were measured for porosity in the laboratory. They revealed values ranging from 0.230 to 0.353 and bracket all but one of the values determined by surface gravity measurements. On the basis of this comparison, the gravity method proposed here is suggested as a workable method that could serve as a valuable reconnaissance tool in estimating the porosity of unconsolidated sediments in certain geological settings.

An example for the intermontane basin model. A variant of the method, applicable to intermontane basin deposits of greater lateral extent, is illustrated in Figure 12. Figure 12a, modified slightly from that of Zbur (1963, Fig. 22), shows values of residual Bouguer gravity at selected stations in Indian Wells Valley, California (Figure 2), plotted against seismic depths determined at the same locations. Included are lines showing the theoretical gravitational attraction of infinitely extending horizontal plates having density contrasts with the bedrock as indicated. Zbur (p. 74) pointed out that infinite plate approximations are not entirely valid for layers at the base of the section. However, corrections to assure greater accuracy could be made if necessary by treating the lower layers as plates of finite extent, provided their horizontal dimensions had been determined seismically. A significant fact apparent from Figure 12a is that the average density contrast of the sedimentary section in Indian Wells Valley decreases as its total thickness increases. This is because of a downward increase in the bulk density of the fill. For a section 2000 feet thick, the *average* density contrast between fill and basement rock is 0.45 gm/cc, while for a section 5000 feet thick, this contrast is less than 0.40 gm/cc.

Contents

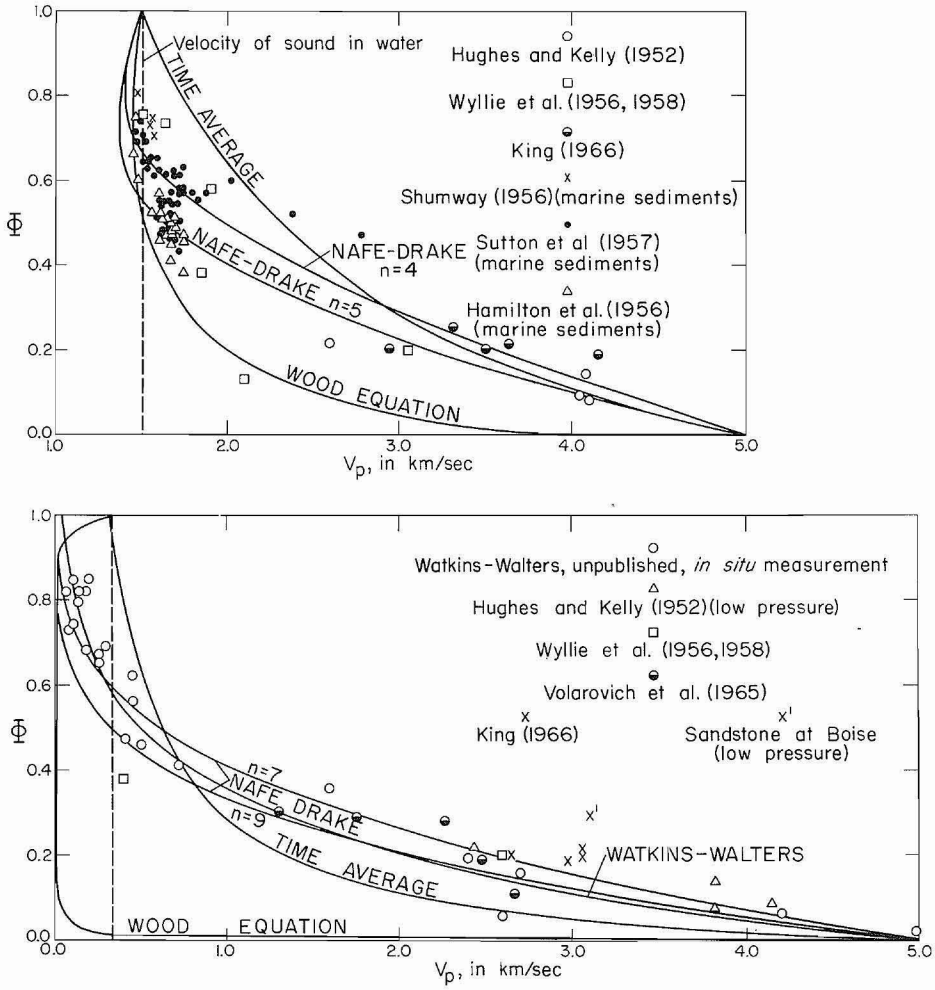


Figure 13. Compressional wave velocities in saturated and unsaturated rocks. Curves represent theoretical expressions relating velocity and porosity. (a) Wave velocities observed in saturated rocks or sediments, under low confining pressures, as a function of porosity. Values assumed in calculation of the theoretical curves: $V_{P_{grain}}$ = 5.0 km/sec; $V_{P_{fluid}}$ = 1.5 km/sec. (b) Wave velocities observed in unsaturated rocks and sediments, under low or negligible pressures, as a function of porosity. Values assumed in calculation of the theoretical curves: $V_{P_{grain}}$ = 5.0 km/sec; $V_{P_{fluid}}$ = 0.33 km/sec; and q = 0.6 (wood equation only).

If the average bedrock density in the area were known, values for density contrast between bedrock and fill could be converted algebraically to fill densities and these, in turn, to porosities, presuming grain density data were available. Von Huene (1960) cites a figure of 2.67 gm/cc for the average bedrock density in Indian Wells Valley. Assuming a grain density of 2.67 gm/cc, and knowing that the water-table is within 100 feet or less of the surface over much of the valley (Zbur, 1963, p. 65), it is then possible to convert values for density contrast into values for porosity. The observational data of Figure 12a have therefore been replotted in Figure 12b, with a smoothed curve fitted to them by inspection. The straight lines in 12b represent the theoretical gravitational attraction of infinitely extending horizontal plates with porosities as indicated. This diagram may be interpreted in the same manner as that of 12a. Thus the average porosity of a sedimentary section 2000 feet thick is approximately 0.285 and that of a section 5000 feet thick, 0.240. Because these averages represent cumulative, weighted values that include the uppermost, porous sediments of the section in each case, it is more useful to compare the *slope* of the observed gravity anomaly vs. seismic depth curve with the slopes of the lines representing porosity. Breaking the curve down into increments of 1000 feet, one can estimate average porosities for the sediments in Indian Wells Valley layer by layer. The results are shown in Table VI. Zbur (p. 63) reported surface sediment porosities of 0.35, but no other data are available from the area for checking the values in Table VI.

The possibility of estimating porosity by the seismic method. It was noted earlier in this paper that there is a pronounced correlation between porosity and compressional wave velocity in clastic rocks, with velocity increasing as porosity decreases. The remainder of this section of the paper is devoted to a brief review

Table VI. Porosity estimates for unconsolidated sediments in Indian Wells Valley, California, based on gravity and seismic data by Von Huene (1960) and Zbur (1963).

Depth below Surface		Approximate Porosity
(meters)	(feet)	
0 - 305	0 - 1,000	0.30
305 - 610	1,000 - 2,000	0.27
610 - 915	2,000 - 3,000	0.24
915 - 1,220	3,000 - 4,000	0.21
1,220 - 1,525	4,000 - 5,000	0.18
1,525 - 1,830	5,000 - 6,000	0.15

Table VII. Summary of investigations into factors affecting seismic wave velocities in rocks and unconsolidated sediments.

Reference	Principal Relationships Investigated	Experimental Method	Lithology	Conclusions
Hughes and Kelly (1952)	Variation of compressional wave velocity with degree of saturation and confining pressure	Laboratory transmission of timed pulse through cores 5 - 10 cm. long	Sandstone and sandy shale	The effect of water content depends on confining pressure. At low pressures (50 bars) velocity increases rapidly with small saturations (0.0 to .10), remains constant in the saturation range from .10 to .90 and decreases in the range from .90 to 1.00. At 50 bars velocity of saturated specimens is greater than that of dry specimens. At higher confining pressures, increase in velocity at low saturation decreases and at 500 bars disappears. Velocity is nearly constant up to saturations of .90 and decreases rapidly as complete saturation is approached. *Note:* All specimens were jacketed and pore fluid pressure was not monitored nor varied independently.
Hamilton, Shumway, Menard and Shipek (1956)	Interrelationships between compressional wave velocity, porosity, density, and grain size	a) *In situ* (sea bottom) transmission of timed pulse between transducers 0.7 meter apart b) Laboratory resonance of sediments in cylindrical container	Unconsolidated clastic marine sediments	Velocity increases as porosity decreases. Porosity increases as grain size decreases. Velocity increases as grain size increases.
Shumway (1956)	Variation of compressional wave velocity and attenuation with grain size	Laboratory resonance of sediments in cylindrical container	Unconsolidated clastic marine sediments	Velocity increases as grain size increases. Attenuation increases as grain size increases.
Wyllie, Gregory and Gardner (1956)	Variation of compressional wave velocity with porosity, pore fluid composition, degree of saturation and lithology	a) Laboratory transmission of timed pulse through columns of synthetic aggregates	Synthetic aggregates	Velocity increases as porosity decreases. Velocity of aggregates of spheres of uniform size is independent of size. Velocity of dry aggregates is less than that of saturated aggregates, but varies with composition of the saturant.
		b) Laboratory transmission of timed pulse through cores 2.0 and 4.5 inches long	Sandstone	Velocity of saturated sandstone is greater than that of dry sandstone. Velocity variations (at 1 atm) are nonsystematic at saturations below .10; between .10 and .70 saturation, velocity is essentially constant; and at saturations greater than .70, velocity increases markedly with increase in saturation. Velocity increases with increasing confining pressure at saturations of 0.0 and .95, but is nearly constant at a saturation of 1.00.
		c) *In situ* logging of interval velocity in wells.	Sandstone and limestone	Velocity increases as porosity decreases.
Nafe and Drake (1957)	Variation of compressional wave velocity with depth; derivation of velocity-porosity and velocity-density functions	Seismic refraction measurements in shallow and deep water marine sediments	Unconsolidated and semiconsolidated clastic sediments	Velocity increases as porosity decreases and bulk density increases. Velocity increases with depth in the sedimentary column.

(*continued on next page*)

of the present state of knowledge concerning factors influencing seismic wave velocities in porous media, particularly as they may apply to unconsolidated aquifers. Table VII presents data from nine investigations of factors affecting seismic wave velocities in natural rocks and unconsolidated sediments, or in artificial rock analogs. The conclusions of the investigators are summarized qualitatively in the last column of the table and although there is considerable repetition in certain of the conclusions, there are also important differences, some of which stem from differences in the simulated or natural environments of the materials during study.

Semiempirical or empirical mathematical models have been proposed by some of the investigators for explaining the observed relationships between velocity and other physical properties. Velocity-porosity equations based on these models are illustrated in Figure 13, where they are shown for comparison with observed data. For a summary and discussion of these models, the reader is referred to Sutton, *et al.* (1957, p. 801 - 811).

The effect of porosity on compressional wave velocity. Figure 13a shows a plot of compressional wave velocities versus porosities for saturated rocks reported in the papers cited in

Table VII. Summary of investigations into factors affecting seismic wave velocities in rocks and unconsolidated sediments. (concl.)

Reference	Principal Relationships Investigated	Experimental Method	Lithology	Conclusions
Sutton, Berckhemer and Nafe (1957)	Variation of compressional wave velocity and attenuation with porosity, grain size, sorting and carbonate content	Laboratory transmission of timed pulse through cores of marine sediments 5.4 cm long	Unconsolidated argillaceous and calcareous lutites, siltites and arenites	Velocity increases as porosity decreases. Velocity increases as grain size increases. Velocity increases as carbonate content increases. Degree of sorting has no observable effect on velocity except that unexplained variations in velocity are greater for more poorly sorted sediments. Attenuation is greatest in coarse-grained, high-velocity sediments.
Wyllie, Gregory and Gardner (1958)	Variation of compressional wave velocity with external pressure, differential pressure (i.e., confining pressure minus pore pressure), porosity pore fluid composition and degree of saturation	Laboratory transmission of timed pulse through cores of unspecified length	Sandstone, dolomite, radiolarian earth, diatomaceous earth and synthetic aggregates	Velocity increases with increasing external pressure. Velocity increases with increasing differential pressure; this increase is marked at low differential pressures, but decreases with increasing differential pressure until an approximately constant, limiting velocity is attained. Limiting velocity increases as porosity decreases. Limiting velocities are not readily achieved in unconsolidated materials, even at high differential pressures. Velocity in sandstone is a function of the saturant velocity when the latter exceeds that of water. For zero differential pressures, velocity increases markedly with external pressure at low values of external pressure, but at higher values, the rate of increase diminishes. At constant uniaxial pressure and with the pore fluid pressure at 1 atm. Velocity increases very slightly with increasing saturation. Velocity in rocks having vugular, as opposed to intergranular, porosity, is not a systematic function of total porosity.
Peselnick (1962)	Variation of compressional and shear-wave velocities with density in dry and saturated carbonate rock	Laboratory transmission of timed pulse through cores 0.75 to 2.0 inches long	Limestone	Both compressional and shear-wave velocities in dry limestone are linear functions of density at 1 atm and 25°C, increasing as density increases (or porosity decreases). Water saturation of limestone decreases both compressional and shear-wave velocities.
King (1966)	Variation of compressional and shear-wave group velocities with hydrostatic confining pressure, differential pressure (i.e., confining pressure minus pore fluid pressure), and pore fluid composition	Laboratory transmission of timed pulses through cores up to 1.75 inches long	Sandstone	Velocity increases as hydrostatic confining pressure increases. Compressional wave velocities are consistently higher for saturated rocks than for dry rocks, the difference being greater at low confining pressures. Shear-wave velocities are generally higher for dry rocks than for saturated rocks at high confining pressures, but at low confining pressures this is true for only one of the rocks investigated, the others showing higher shear-wave velocities for kerosene saturation than for the dry state. For saturated rocks, both the compressional and shear-wave group velocities are nearly constant for equal increases in confining and pore fluid pressure (i.e., constant differential pressure, with increasing confining pressure). Shear-wave amplitudes depend much more on confining pressure and composition of the pore fluid than do compressional wave amplitudes.

Table VII. Curves for the time-average equation and Wood equation (Wyllie, *et al.*, 1956) and Nafe-Drake equation (Nafe and Drake, 1957), for $n = 4$ and 5, are shown for comparison with the observed data.* The time-average equation fits well for low porosities, the Wood equation fits well for high porosities and the Nafe-Drake equations, for all of the data.

Figure 13b shows compressional wave velocities versus porosities for unsaturated rocks. Included with data from the previously mentioned investigations are unpublished data and a derived empirical equation from a study by J.S. Watkins and L.A. Walters of the U.S. Geological Survey. The time-average equation, Wood equation, Nafe-Drake equation (for $n = 7$ and 9) and the Watkins-Walters equation are shown for comparison. The Nafe-Drake equation was evaluated for $n = 6,7,8$ and 9. The curve for $n = 7$ fits the data well for porosities less than 0.7. For porosities between 0.7 and 0.85, it tends to predict values lower than those observed.

For practical purposes, the difference in fit between the Nafe-Drake equation for $n = 7,8$ and 9 and the Watkins-Walters empirical equation is not significant except at very high porosities

*The quantity n is an arbitrary exponent in the Nafe-Drake equation.

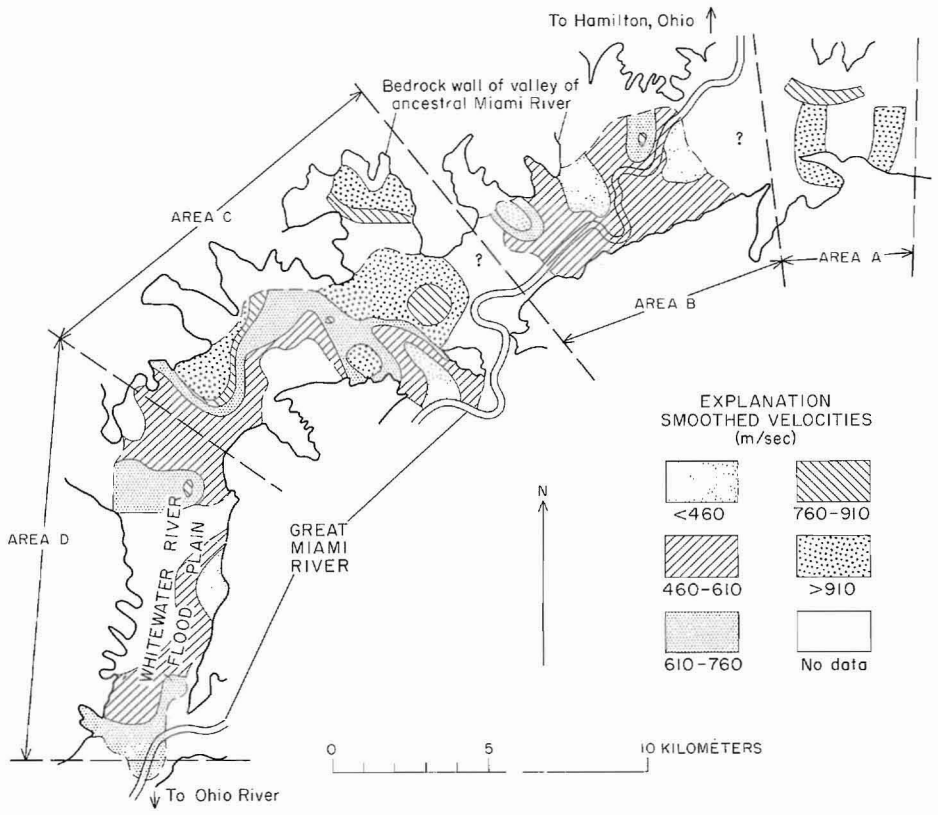


Figure 14. Map of smoothed compressional wave velocities observed in unsaturated alluvium of the ancestral Miami River Valley between Hamilton, Ohio, and the Ohio River.

where velocities from the latter expression best represent the porosity data. From the results illustrated in Figure 13, average errors of less than 15 percent in the estimation of porosities from seismic velocity data apparently are possible, even though the data were taken from a number of sources. By restricting data to certain sediment types or certain areas, significantly better estimates of porosity could be obtained.

The effect of confining pressure on wave velocity. Confining pressure, which increases with depth in unconsolidated aquifers as a result of increasing overburden weight, has an effect on seismic wave velocity, as well as on density. Investigations have been made of the depth dependence of velocity and density in many settings and Woollard (1962, Figs. 1, 2, 6, 12 and 13) has presented a graphical summary of the results for Recent sediments. Velocity increases and porosity decreases with increasing depth and confining pressure.

If a pore fluid is present in the sediments, its effect on seismic wave velocity will depend largely on whether the pore pressure varies independently of that of the rock frame. In saturated impermeable rocks, the pore pressure may nearly equal the confining pressure, yielding very low differential pressures. In highly permeable sediments, large differential pressures may exist. King (1966) found from laboratory experiments that seismic wave velocities in saturated porous rocks depend very simply on the algebraic difference between static confining pressure and internal pore pressure. Thus for a constant confining pressure, an increase in pore pressure will lead to a reduction in velocity.

Extrapolating these findings to a natural setting, this observation could be of importance in some of the large intermontane basins of the western United States, for the sedimentary fill in some of these basins includes more than one aquifer. Commonly, several aquifers are separated from one another by aquicludes that block hydraulic continuity between them. Hence pore pressure magnitudes may vary discontinuously from stratum to stratum, even though confining pressure increases systematically downward. Attempts to estimate porosities from seismic velocity data under these circumstances could lead to larger errors than might be expected on the basis of data shown in Figure 13.

The effect of saturation on wave velocity. King (1966) found that although compressional wave velocities are consistently higher for saturated rocks than for dry rocks, especially at low confining pressures, shear-wave velocities are generally higher for dry rocks. This suggests a practical hydrogeological application for measurement of shear-wave velocity with compressional wave velocity.

Molotova and Vassil'ev (1960) reviewed available data regarding the ratio of observed compressional to shear-wave velocities. They found that in sedimentary materials this ratio varies from 1.5 to 14.0, with values greater than 2.0 being restricted to rocks with compressional wave velocities between 1.0 and 2.0 km/sec (3300 and 6600 ft/sec). Large values of this ratio are further limited to materials within approximately 100 meters (330 feet) of the surface. These ranges of depth and compressional wave velocity imply that anomalously low shear-wave velocities are restricted to saturated unconsolidated media.

The effect of grain size on wave velocity. Grain size appears to have an effect on seismic wave velocities in sediments, although the observational data are as yet meager. Sutton, *et al.* (1957) observed that compressional wave velocity increases linearly with the logarithm of median grain diameter for sea floor sediments in the size range 0.002 to 0.27 mm. High correlation coefficients

indicated that this velocity - grain size relationship was independent of porosity. The relationship is probably applicable to fine-grained fluvial, lacustrine and glacial sediments, but full confidence in it requires a study of additional data. In particular, the effects of sorting on velocity over a wider range of sorting than that considered by Sutton, *et al.* needs further investigation.

The effect of mineralogy on wave velocity. Almost all investigations of seismic-wave velocity have shown a strong velocity dependence on the lithologic composition of the rocks. Rocks of different mineral composition have different elastic moduli. Too few studies have been made of unconsolidated sediments to indicate the relative importance of this factor, particularly in relation to the effects of grain geometry and grain contact. The study of Sutton, *et al.* (1957) indicates that for fine-grained, uniform sediments, carbonate content is important. Whether this observation is valid for coarser, more poorly sorted materials remains to be seen. It could be of importance in studies of unconsolidated aquifers in which the sediments were derived from source areas of different composition or where some stratigraphic units in the section have been cemented, or partially cemented, and others have not. Interaction between clay mineral grains and the water contained in the interstices of the sediment may also affect velocity, but this phenomenon has not been studied adequately and the magnitude of the effect is therefore not known.

The effect of cementation on wave velocity. The effect of cementation can be seen by comparing velocities for consolidated and unconsolidated materials. Published (Table VII) and unpublished data of Watkins and Walters show that for saturated *unconsolidated* sediments, compressional wave velocities range from 1.46 km/sec (4800 ft/sec) to 4.20 km/sec (13,800 ft/sec), with the preponderance of values between 1.46 and 1.80 km/sec (4800 and 5900 ft/sec). Velocities of saturated *consolidated* rocks range upward from 2.25 km/sec (7400 ft/sec). The upper limits of saturated and unsaturated rock velocities converge to 5.0 or 8.0 km/sec (16,400 or 26,200 ft/sec), depending on the mineral composition of the rock. The velocity range between 1.80 and 3.00 km/sec (5900 and 9800 ft/sec) (Figure 13a) is characteristic of weakly cemented, saturated rocks.

Velocities of unsaturated, unconsolidated sediments and soils range from 0.08 km/sec (260 ft/sec), to 0.73 km/sec (2400 ft/sec). Most values lie between 0.3 and 0.5 km/sec (1000 and 1600 ft/sec). Velocities of unsaturated, consolidated rocks range from 1.30 to 5.00 km/sec (4300 to 16,400 ft/sec), with values distributed throughout the range.

The effect of cementation is to increase rigidity and, in most cases, incompressibility, of the rock or sediment, at the same time decreasing porosity. Therefore both shear and compressional wave velocities increase with increasing cementation. Although it is generally not required in groundwater studies, the differentiation of dry, weakly cemented rocks from saturated, unconsolidated sediments with similar compressional wave velocities could probably be accomplished by measurement of shear-wave velocities which would be higher in dry, partially cemented rocks.

In summary, seismic-wave velocities in unconsolidated or semiconsolidated sediments vary with porosity, confining pressure, differential pressure, degree of saturation, grain size, degree of cementation and possibly mineralogy. However, porosity is the property that correlates best with velocity, offering hope that velocity values can be used for estimates of porosity in unconsolidated aquifers. For unsaturated sediments, the Nafe-Drake equation for n = 7,8 or 9 expresses the relationship very well for porosities of less than 0.70, but yields velocities lower than those observed for porosities greater than 0.70. For saturated sediments, the Nafe-Drake equation for n = 4 or 5 expresses the porosity-velocity relationship best. *In situ* measurements of both velocity and porosity in typical, near-surface, unconsolidated, continental sediments are very much needed, if the hydrogeologist is to use seismic measurements to elicit information on water-bearing properties of the sediments. This is an area of investigation in need of further research.

An example for the buried valley model. To illustrate the importance of the potential of velocity measurements in estimating porosity, the reader's attention is directed to the results of two field investigations by the writers. Both investigations illustrate lateral variations of the compressional wave velocity in unconsolidated sediments.

Figure 14 shows the distribution of smoothed compressional wave velocities in unsaturated alluvium associated with sediments of the ancestral Miami River between Dayton, Ohio, and its confluence with the Ohio River. The valley of this ancestral river is filled largely with permeable outwash sand and gravel with thicknesses ranging from 30 to 100 meters (100 to 330 feet). The course of the present Great Miami River follows that of the ancestral river for much of the distance under consideration. Bedrock in the area consists of impermeable Upper Ordovician limestones and shales. The deep fill of permeable glacial debris in the old valleys constitutes the area's main source of groundwater.

Thickness of the glacial debris was investigated in 1962 and 1963 and reported by Watkins and Spieker (1965). During the investigation refraction profiles were spaced closely enough to show general trends in velocities in both unsaturated and saturated alluvium. The observed velocity data could not be readily contoured because of locally abrupt variations and consequently they were smoothed by gridding. A grid with a unit dimension of 0.4 km was established and averages of observed velocity values between 0.0 and 0.4 km and between 0.4 and 0.8 km were calculated at each grid center. The averages for observed velocities within 0.4 km of the grid centers were multiplied by 0.75, and the averages between 0.4 and 0.8 km, by 0.25. The two averages were then summed. The velocity contours shown in Figure 14 represent smoothed velocities for unsaturated alluvium. The nature of variations in the smoothed data for saturated alluvium could not be contoured with confidence, but a general decrease in velocity from northeast to southwest was observed. This probably indicates a higher porosity to the southwest, possibly as a result of decreased grain size and better sorting of the outwash materials.

Figure 14 is divided somewhat arbitrarily into four areas on the basis of the smoothed velocities of unsaturated, unconsolidated alluvium. The lateral variations in velocity shown may be regarded as reflecting, at least in part, lateral variations in porosity. If it were possible to convert the velocity data directly to porosities, seismic measurements of unconsolidated sediments would provide a field technique for pinpointing deposits of a specific porosity.

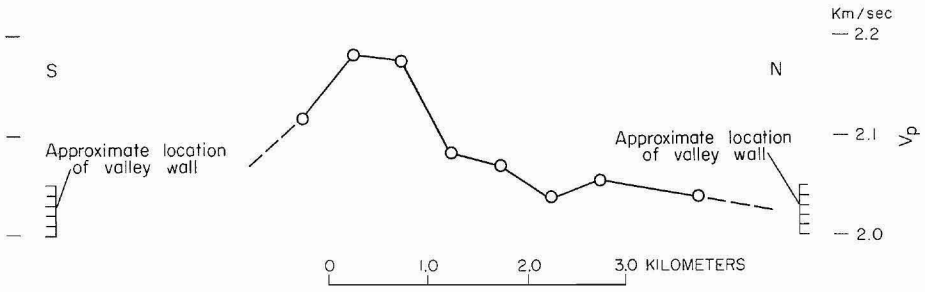


Figure 15. Lateral variation in compressional wave velocities observed in saturated alluvium across Chuckwalla Valley, west of Blythe, California.

An example for the intermontane basin model. Figure 15 shows variations in velocity for saturated alluvium recorded during a seismic refraction traverse of Chuckwalla Valley, California. Velocities in alluvium increase from the north wall southward, reaching a maximum near the center of the basin, then diminishing southward. The average grain size probably decreases toward the center of this valley, as it does in most desert basins, and the average sorting probably improves in that direction also. Improvement in sorting should lead to an increase in porosity and one might expect a decrease in velocity from the margins toward the center of the valley.

The increase in velocity observed suggests that possibly factors other than porosity affect velocity here. One or more of several salts that tend to precipitate readily in sediments of desert environments may be responsible for the increase in velocity in the alluvium of Chuckwalla Valley. Their effects could be due to their inherently high velocities, increased rigidity and incompressibility of the alluvium resulting from cementation by the salts, porosity reduction resulting from the filling of primary pores in the clastic sediments by the salts. Possibly all three factors are involved. An investigation into this matter would shed considerable light on the usefulness of velocity data in estimating porosities in desert basin sediments.*

The data of Figures 14 and 15 both reveal conspicuous lateral velocity variations in unconsolidated sediments that are associated with variations in other physical properties of the sediments. Seismic velocities in such sediments should be investigated intensively for their potential to provide information on water-bearing properties of alluvium.

Areas for further research

Most published work on applications of the seismic and gravity methods to unconsolidated sediments has involved their use to determine layer thickness. Possible means of measuring or estimating porosity, at least approximately, have been suggested here. In the concluding paragraphs, some possible avenues for further research are suggested, the objective being improvement of *in situ* surface geophysical measurements as applied to hydrogeology.

Seismic methods

Attenuation and spectral characteristics. Little is known of the attenuation of seismic energy in water-bearing alluvium and rock except that for compressional waves, it is greater in unsaturated, than in saturated, media. The opposite may be true for shear waves (King, 1966). Berson, *et al.* (1959) have suggested that contrasts in porosity and grain size of aquifer sediments might be recognized from studies of both the frequency spectrum and absorption characteristics of refracted energy. In stratigraphic sections they studied, where definite changes in physical properties were not particularly evidenced by contrasts in velocity, water-bearing sand was characterized by higher frequencies than argillaceous strata. Thus it may be possible to distinguish water-bearing formations from others of similar velocity on this basis. Energy losses and frequency spectra in unconsolidated media therefore may be useful parameters in hydrogeological investigations, but far more data are required to evaluate fully their usefulness.

Shear-wave velocities. King (1966) found that, in general, shear-wave velocities are higher in dry rocks than in saturated ones (the opposite is true for compressional wave velocities), and that shear wave amplitudes are more sensitive to the composition of the pore fluid and confining pressure than are compressional wave velocities. Both these facts suggest that shear-wave investigations of unconsolidated aquifers may be fruitful.

Measurement of compressional wave velocities is straightforward and commonplace; measurement of shear-wave velocities is somewhat more difficult. New methods and instrumentation have been introduced, such as those based on rotational properties of shear waves (Whitcomb, 1966), and these new approaches engender hope that instruments which will pass shear waves but reject compressional waves may ultimately be developed for common field use.

Surface waves. Rayleigh waves in alluvium and sediments have been observed by Korschunow (1955), Bulin and Savarensky (1961), and McEvilly and Stauder (1965). The interpretation of these data is commonly accomplished by iterative fitting of observed data to theoretical curves. A recent paper by Mooney and Bolt (1966) provides master curves for a wide range of models. It should not be difficult to calculate additional curves for hydrogeologically relevant situations. Figure 16 shows an example of a possible use of surface-wave dispersion in an aquifer with uniform compressional wave velocity, but with a shear-wave velocity inversion at depth. In this instance, a surface-wave analysis provides data about a subsurface variation to which compressional wave travel times are insensitive. Such an inversion might be due to the presence of interstitial clay which has reacted with the pore-fluid, reducing the rigidity of the sediment. The presence of this clay might, in turn, be important hydrogeologically, causing low permeability and reduced porosity in the sediment.

*J.D. Fett (personal communication, 1967) has observed a similar increase in velocity with decrease in average grain size in large tailings piles. He attributes this relationship to a higher moisture content resulting from a higher specific retention of the fine-grained sediments.

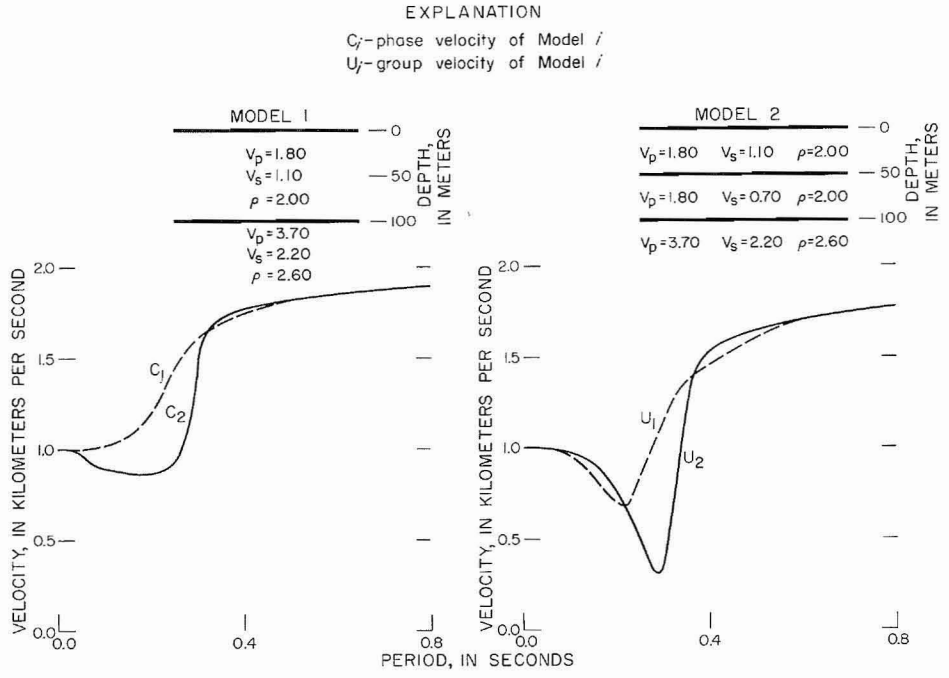


Figure 16. Phase and group velocities calculated for hydrogeological models with two different velocity distributions. Model 2 is similar to Model 1 except for a shear-wave velocity inversion between 50 and 100 m.

Gravity methods. Great strides have been made in the last decade in the field of gravity measurement and interpretation. These have included the development of field instruments with better precision and drift characteristics, development of high-speed machine reduction and model interpretation methods, and improved means, through application of trend surface analysis, of isolating individual parts of an observed gravitational field. Further development of the gravity method in the field of hydrogeology, however, will come not through surface measurements, but through borehole measurements, using an instrument of the type described by McCulloh, *et al.* (1967(a) and (b)) or by Howell, *et al.* (1966). Current logging methods for measuring *in situ* the density of a sedimentary section in a borehole (e.g., gamma-gamma logs) depend on laboratory analyses of core samples for calibration. Use of a borehole gravity meter would allow a more direct measurement of density, in addition to offering the advantage of sensing a much greater volume of rock or sediment. It would be little affected by formation damage from drilling or drilling mud infiltration, and borehole diameter variations would be of considerably less consequence than in the case of gamma-gamma logs. The significance of the borehole gravity meter is discussed in two recent papers by McCulloh (1966, 1967).

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A review of some problems of seismic prospecting for groundwater in surficial deposits

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Abstract. Seismic prospecting is often used in the search for groundwater supplies in surficial deposits. The usual targets in this search are unconsolidated or semiconsolidated deposits of clastic material, confined by less permeable deposits. Besides the normal limitations of the seismic method, three particular problems confront the seismologist. These are: (1) the difficulty of generating simple seismic bodily waves in unconsolidated materials; (2) the lack of a definitive relationship between seismic velocity and a particular deposit; and (3) the much greater significance of errors in small-scale seismic investigation.

These problems are reviewed and their importance in investigating surficial deposits examined. The value of uphole wavefront diagrams in determining the velocity model *in situ* is illustrated by a series of examples. Problems of using the seismic method to map the water-table are considered briefly. Amplitude and frequency studies are promising in removing some of the restrictions imposed by the lack of velocity contrasts. In one example such studies provided useful information on the identity and porosity of an aquiferous sand in the U.S.S.R. However, an attempt to repeat this experiment under similar conditions in Canada failed, probably because of unavoidable changes in shothole characteristics. Finally, the possible role of the electroseismic effect in determining permeability is mentioned.

Résumé. La prospection séismique est souvent utilisée pour la recherche d'eau souterraine dans les formations superficielles. Les objectifs de ce genre de recherche sont les dépôts meubles ou semi-consolidés de matériaux clastiques limités par des sédiments moins perméables. En plus des limites normales de la méthode séismique, trois problèmes viennent particulièrement se poser au séismologue. Ce sont: (1) la difficulté de produire des ondes séismiques dans des matériaux friables; (2) l'absence de relation définie entre la vitesse de propagation et un gisement particulier; (3) la grandeur des erreurs dans les études séismiques à petite échelle.

L'auteur aborde l'étude de ces problèmes et de leur importance dans la prospection des sédiments de surface examinés. Une série d'exemples vient illustrer l'intérêt que représentent les diagrammes de fronts d'ondes verticales dans la détermination des vitesses *in situ*. Les problèmes que pose l'utilisation de la méthode séismique pour déterminer le niveau hydrostatique font l'objet d'une étude rapide. Les études d'amplitude et de fréquence permettent bien, ainsi que cela était prévu, d'éliminer certaines des restrictions attribuables au manque de contraste. L'auteur montre par un exemple que des études de ce genre ont permis d'obtenir des renseignements précieux sur l'identité et la porosité d'un sable aquifère en U.R.S.S. Cependant, un essai de réédition de cette expérience dans des conditions semblables au Canada s'est soldé par un échec, probablement en raison de changements inévitables dans les caractéristiques des trous de tir. Enfin, il est fait mention du rôle possible de l'effet électroséismique dans la détermination de la perméabilité.

Groundwater supplies are often sought in the surficial deposits that lie above the bedrock surface. The most common targets are permeable deposits of unconsolidated, or semiconsolidated, clastic rocks. Water-bearing units of this kind range from many square miles in area, and a hundred feet or so in thickness (e.g., glacial outwash deposits of sand and gravel) to a few tens of square feet in area, and a foot or so in thickness (e.g., alluvial deposits of sand from present-day streams). Delineation of these units is possible by surface mapping for deposits that outcrop or have some form of surface expression (e.g., buried beach ridges). However, many potential aquifers are concealed below other deposits and must be located by a subsurface exploration program.

Drilling is the basic element of a subsurface exploration program. From borehole logs, the nature, depth and thickness of the various deposits can be determined. Subsurface cross-sections or contour maps are then constructed, considering the geological habit of the area. Unfortunately, the cost of drilling is often high and complete exploration of deposits by this means alone is too expensive. One way to reduce this cost is to drill fewer holes and use a geophysical method to provide information over a larger area. Relationships can sometimes be established between geophysical measurements and significant hydrological parameters, such as the water-table and porosity. Under these circumstances, adding one or more geophysical methods to the exploration program is justified, both economically and technically. The aim of this paper is to discuss some of the problems involved in the use of the seismic method for these purposes.

The literature dealing with the subject of seismic prospecting in surficial deposits has been restricted mainly to the presentation of successful case histories. Many of these describe investigations in areas where there are large velocity contrasts between materials of interest, and where fairly simple geological conditions prevail. In areas of low velocity contrasts and more complicated geological conditions, the interpretation of seismic data meets many problems, and it is difficult to decide whether the seismic method should be used. A consideration of the problems involved may therefore be helpful in making this decision.

One or more of the following three groups of problems may impose major restrictions on the use of seismic prospecting in surficial deposits: (1) the difficulty of generating simple seismic bodily waves in the unconsolidated materials; (2) the lack of definitive relationship between seismic parameters and specific types of surficial deposits; and (3) the increased significance of errors in small-scale seismic investigations. The first two will be discussed in the following sections of this paper. The problems introduced by the small scale of shallow seismic investigations have been reviewed previously by Evison (1952) and Domzalski (1956) and will only be considered briefly in the present paper.

The generation of seismic waves in unconsolidated material

A primary requirement for seismic prospecting is the generation, detection and recording of distinct seismic waves. Most interpretation methods are based on travel times of compressional and shear waves. It is therefore imperative that enough of the source energy be converted into these types of waves. In unconsolidated materials, the generation of distinct types of bodily waves is often difficult, if not impossible, and the lack of such waves may constitute a serious problem in the seismic exploration of surficial deposits.

An interesting study of the above problem was made by Korschunow (1955), for various types of loose materials. Typical locations representing the three main classes of soils (bonding material, non-bonding material and soils mixed with organic matter) were chosen for investigation. For each class of soil, Korschunow established that most of the energy was carried by wave groups with a Rayleigh wave mechanism, both from hammer blows on the surface and explosions in boreholes. The compressional wave inside the loose surface layer was found to be coupled to this Rayleigh wave mechanism. A weak head wave, associated with the boundary plane between the surface layer and the first, more consolidated layer below, was the only independent compressional wave recorded.

Korschunow failed in his attempts to increase the depth of penetration of the seismic energy. No waves were recorded from boundaries below the first consolidated layer, in spite of the explosive charge being increased from a few hundred grams to several kilograms. He concluded that the lower boundary of the Rayleigh wave mechanism forms an energy threshold, which can only be overcome by an extremely large increase in explosive charge. This conclusion may explain the lack of identifiable events in some areas where distinct velocity boundaries are known to exist, but where the source of seismic energy has been hammer blows on the surface of the ground, or shooting in the upper unconsolidated layers.

Korschunow's results have been confirmed in many other studies. For example, in a study of explosion-generated seismic pulses on the surface of a sand and pebble layer, Howell and Budenstein (1955) recognized three separate pulses on seismic records at distances ranging between 10 and 1172 feet. At distances of 10 to 180 feet, two pulses were identified as a complex type of compressional wave and as a coupled surface wave, respectively, the latter being similar to the type described by Leet (1946). The up-away, down-toward particle motion indicates that the first pulse is a direct compressional wave, but there is also a strong transverse component of motion. At and beyond 180 feet, a third pulse was identified as the arrival of overlapping Rayleigh and Love waves.

Relative amounts of energy in the three types of pulses were also calculated by Howell and Budenstein from shots of about 500 grams of dynamite. The energy in the first pulse decreased with distance at a much greater rate than in the other two pulses and became negligible, in comparison, at a distance greater than 450 feet from the shotpoint. As the distance was increased even further, energy in the third pulse became predominant and records were characterized by typical motions associated with surface waves.

The strong transverse component associated with the direct compressional wave in the above study has also been noted in other investigations of surficial deposits (e.g., Dobrin, *et al.*, 1954; Burke, 1968). A probable explanation of this effect is that a transverse component is introduced by inhomogeneities close to the shotpoint, or in the material through which the wave is travelling. It has been shown theoretically that for certain kinds of aeolotropic and inhomogeneous media, the compressional and rotational components may be coupled in a complicated manner, and the resultant motion does not separate distinctly into compressional and shear waves (Grant and West, 1965).

In areas where the near-surface layers have velocities less than the velocity of sound in air, ground-coupled air waves are another potential source of misinterpretation. Mooney and Kaasa (1962) demonstrated the existence of this type of wave in recordings made from hammer blows. In an area of thick and fairly uniform soil, ground-coupled air waves occurred as low-amplitude, high-frequency (40 to 500 cps) first arrivals, with a much lower attenuation rate than the later-arriving direct wave through the soil. It was suggested that careful control of the recorder amplification, or filtering of the high-frequency components, might eliminate the ground-coupled air waves. Unfortunately, this can have a deleterious effect on the recording of direct and refracted waves as the distance from the energy source is increased.

Significant errors may be introduced if a ground-coupled air wave is mistakenly identified as a direct wave through the upper layer. This is especially true when the velocity of compressional waves in the upper layer is much less than the velocity of sound in air, and the thickness of this layer is large compared to the depth of interest. Mooney and Kaasa (1962) give a table of errors based on typical velocity values. They show that the error becomes greater than 100% when the upper layer velocity is 600 feet/second and the lower layer velocity is 2000 feet/second. Near-surface velocities that are favourable to the generation of ground-coupled air waves occur in many places, particularly where the topsoil is derived from underlying surficial deposits (e.g., Hobson and Collett, 1960). However, the ground-coupled air wave has not been a problem in Canada at source-to-detector distances in excess of 20 feet using a hammer source and, therefore, only constitutes a problem in very shallow investigations (Hobson, written communication, 1968).

The foregoing discussion shows that the simple direct reading of a time of arrival from a record and its blind association with a specific seismic event causes misleading interpretation. Incorrect depth estimations are possible and erroneous identification of deposits also may result if velocity alone is used as a criterion. This is particularly true if only the time of arrival of the first event is obtained, and no record of particle motion, as with the many instruments designed on a time-counting principle. Even with those instruments capable of producing a trace recording of first and secondary arrivals, events chosen on the basis of 'character change' or an 'appropriate time of arrival' may yield nothing more than a phase velocity for part of a surface-wave mechanism. According to an abstract of a recent paper (Geophysical Abstracts, 1967), Millon has reported that shear and surface waves have been mistakenly identified as low-velocity compressional waves in hammer surveys. This misidentification led to errors in interpretation and low estimates of elastic parameters of the ground.

Determining the propagation mechanism for a particular seismic event will do much to eliminate the ambiguity in identifying different waves. If misleading results are being obtained, three-component recording or uphole wavefront diagrams (see next section) should be used to determine the real identity of the event. In a completely unknown area, it is good practice to include such studies as an integral part of the preliminary investigations.

The relationship between seismic parameters and surficial deposits

Seismic velocity. The interpretation of seismic records is based on the correlation of seismic events with subsurface boundaries. In most surveys this means that there should be a difference in seismic velocity between deposits. It is therefore helpful to consider the factors which control the velocity in such media.

A major obstacle in assigning a representative velocity to a surficial deposit might appear to be the latter's heterogeneous composition. However, it must be remembered that homogeneity is a relative, rather than an absolute, characteristic of matter, dependent on the scale of the problem under consideration. In seismic wave transmission, the determining factor has been found to be the shortest wavelength. As long as the properties which determine velocity can be considered the same when averaged through a volume, small compared to the shortest wavelength of interest, the medium may be considered as grossly homogeneous. There is an extensive literature on seismic wave propagation in grossly homogeneous media and only a brief review of the major results will be given here.

For a medium, which is a two-component mixture of a solid and fluid, two sets of conditions have generally been considered: (a) a medium represented by a liquid suspension of solid particles without rigidity; and (b) a medium represented by a framework of solid particles in elastic contact, with the pore spaces completely saturated with the fluid.

For the first set of conditions, the equation for the velocity of the compressional wave has been given by Heiland (1940). The compressional wave velocity is found to be a function of the volume-weighted average of the densities and effective bulk modulus of the two constituents, the values of which are both dependent on the porosity of the mixture. Extremely low velocities, less than the compressional velocity of either constituent, are possible in water saturated sediments of high porosity (Officer, 1955). Agreement is close between calculated velocities and those measured in loosely compacted, water-saturated, marine sediments, although empirical modification of the relationship is necessary in some cases (Nafe and Drake, 1957). However, most continental surficial deposits have some rigidity and it is more appropriate to consider models which allow elastic contact between the particles.

Several workers have used a model of an unconsolidated granular medium, with a regular packing of elastic spheres confined under its own weight, as a basis for the second set of conditions. Gassmann (1951) chose a model with an hexagonal packing of uniform spheres, the latter being deformed at their points of contact by the pressure of the overburden. He assumed that the elastic moduli of such a system could be determined from static deformation of the completely unsaturated medium. Relationships were derived for the velocity of both compressional and shear waves in a completely dry and a fully saturated medium. The results show that the velocity, besides being a function of the elastic parameters, density and porosity, is also proportional to the sixth root of the depth of burial in the medium.

White and Sengbush (1953) developed a similar theory for the cubic packing of uniform spheres, but allowed for tangential stresses at the point of contact of the spheres. Although a cubic packing lacks mechanical stability and is not a very realistic model, it does illustrate the more important features of a regular packing of spheres. With this model, the velocities of compressional and shear waves in a completely dry medium were again found to be dependent upon the sixth root of the depth of burial. The velocities approach a value of zero towards the free surface. White and Sengbush verified this dependence of both velocities on the sixth root of the depth of burial by borehole measurements in naturally occurring, clean sands. An expression equivalent to that of Gassmann was obtained for the compressional wave velocity in a fully saturated medium. A rapid increase in compressional velocity, as the sand became fully saturated with groundwater, was measured at a depth of 50 feet in one of the boreholes.

Brandt (1955) has derived a series of expressions for the velocity of compressional and shear waves in a liquid-saturated packing of randomly stacked spheres of different sizes. Relationships similar to those of Gassmann and White and Sengbush are obtained for this model, if the latter is restricted to simple arrangements of spheres (Hicks and Berry, 1956). Brandt also showed that the velocity would be a function of $(p_e - np_i)$, where p_e is the external confining pressure, p_i is the internal pore pressure and n ranges between 0 and 1. Laboratory measurements on sandstone samples have shown that n has a value of 1, if the effects of mechanical hysteresis are avoided in the loading and unloading process (e.g., Gardner, *et al.*, 1965; King, 1966).

In the models referred to, the contained fluid and framework were assumed to move together and to each have the same displacement in any elementary volume. A more fundamental treatment of the whole problem has been made by Biot (1956a, 1956b, 1962), in a series of papers in which the movement of the fluid relative to the framework was considered. The existence of two coupled compressional waves and one shear wave was predicted for the case of a fluid-saturated, statistically isotropic, porous medium. At low frequencies, viscous effects are dominant and only the first kind of compressional wave, with fluid and solid moving together, is significant. Low-frequency representatives of the second kind of compressional wave have extremely small velocities and are highly attenuated because of their diffusiveness. At higher frequencies, inertial effects become significant and the second kind of wave is also propagated through the medium for some distance. Paterson (1956) has identified this latter type of wave in laboratory measurements made on a Pleistocene beach sand at ultrasonic frequencies. However, the critical frequency at which the transition between the two types of behaviour occurs, is well above the frequencies normally encountered in seismic investigations of surficial deposits.

Near-surface deposits are often not fully saturated with water. A discussion of the factors controlling velocity would therefore not be complete without considering the influence of the degree of saturation. Again several approaches of varying complexity are possible depending on the selection of the model.

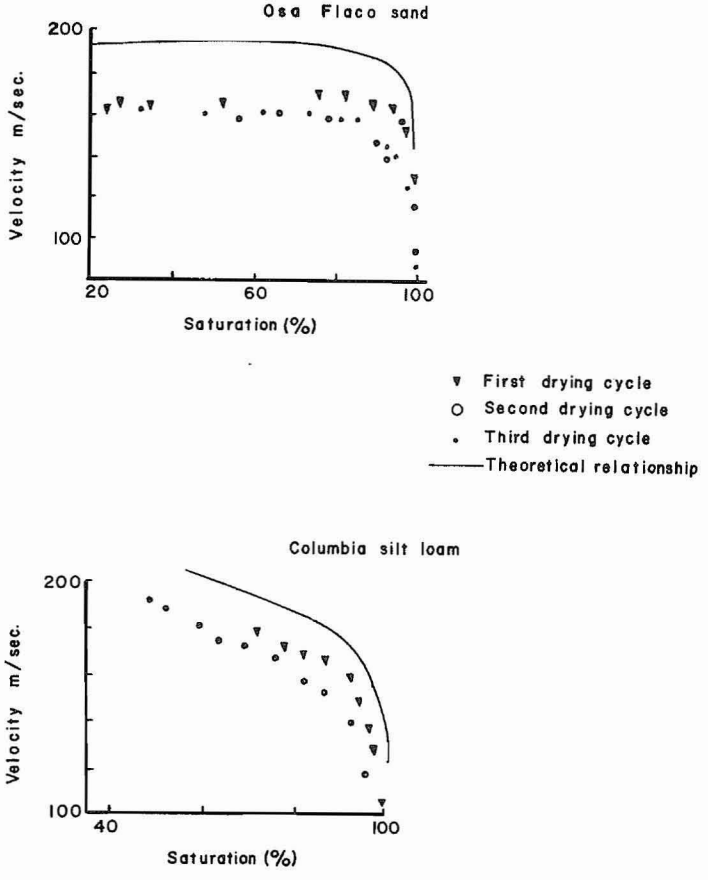


Figure 1. Variation of velocity with saturation (after Brutsaert and Luthin, 1964).

Berson, *et al.* (1959) have examined two of the simpler models. First, they used a similar approach to Heiland (1940), except that a three-component mixture of air, liquid and solid was taken as a model. The compressional velocity within this mixture was found to be a function of the ratios of the densities, velocities and volumes of the three components. Velocities in sands of different porosities were calculated for varying degrees of saturation. The results all show that the velocity remains near to the value in a completely air-filled sand, until almost 100% saturation is reached. The velocity then rises abruptly to the value for a fully saturated, two-component mixture. The second model considered by Berson, *et al.* was that of a cubic packing of spheres, compressed by its own weight. Similar results to those above were reported for this model, with the compressional velocity varying only slightly with partial saturation and suddenly increasing in the vicinity of 100% saturation.

Brutsaert (1964) has developed a general theory for waves propagating in an unconsolidated medium containing both a gas and a wetting liquid, using a similar procedure to Biot. The elastic constants were derived using the model proposed by Brandt (1955), but modified to allow both gas and liquid in the pore spaces to be in diffusive contact with the atmosphere. Three types of compressional wave and one type of shear wave were predicted by this theory. At low frequencies, only the non-dissipative, normal type of compressional wave was found to be of importance.

The abrupt increase of compressional velocity at full saturation may be reversed under conditions of low overburden pressure. In laboratory measurements on a sand and a loam (Figure 1), Brutsaert and Luthin (1964) found a rapid increase in velocity if a previously fully saturated sample was slightly desaturated. An examination of the relationship between the velocity and the degree of saturation for the same conditions as in the experiments, using Brutsaert's previously developed theory, confirmed the decrease of velocity on desaturation. At low confining pressures, it was established that moisture suction is a predominant factor in determining the velocity. The increase in moisture suction during the desaturation process causes the individual soil particles to be drawn closer together. This compression results in higher values of the appropriate elastic moduli and the over-all effect is an increase in velocity with desaturation.

Several workers have suggested that the abrupt increase in compressional velocity with approach to full saturation might be used to map the water-table, as is stated in many textbooks on hydrology (e.g., Davis and DeWiest, 1966). The success of this application will depend on how close the boundary between the partially and fully saturated zones approximates to 'the surface in unconfined material along which the hydrostatic pressure is equal to the atmospheric pressure', which is the hydrologist's definition of the water-table (Davis and DeWiest). In loosely consolidated, coarse-grain deposits, such as sand, the approximation is generally good. However, in clay and silt deposits, the zone of full saturation may extend in an irregular fashion into the capillary fringe and the boundary between full and partial saturation will probably bear little relationship to the water-table. In addition, if the water-table is near to the surface in a deposit with a high inherent moisture suction, there will be no increase in compressional velocity. At greater depths the contribution of the moisture suction becomes insignificant compared to overburden pressure and the compressional velocity is again a maximum for a fully saturated medium, decreasing with desaturation. The depth of the water-table is therefore a determining factor in the success of the seismic refraction method in such deposits.

An excellent review of the various criteria which may be used in identifying compressional head waves associated with the zone of saturation is given by Levshin (1961). From field studies in the U.S.S.R., he found that these head waves had an apparent velocity ranging between 1500 m/sec and 2500 m/sec and their time-distance plots fitted well to a straight line. At shallow depths, the head waves associated with the water-table had higher predominant frequencies than head waves associated with lithological boundaries. Another useful feature established was the absence of corresponding shear head waves associated with the water-table. In fact, theoretically, the shear-wave velocity in a fully saturated medium should be slightly less than in a completely dry medium (White and Sengbush, 1953). Unfortunately, this is not a unique feature for identifying the water-table because a similar absence of shear waves is noted for some clay deposits.

The foregoing discussion shows that the velocity of a particular type of deposit may be a complex function of many individual factors. The variations of these factors in real media are difficult to predict theoretically and it is usually necessary to obtain velocity control for a survey by empirical means. Because of the low velocity of the surficial deposits, compared to that of the borehole fluid, continuous velocity logging of testholes is

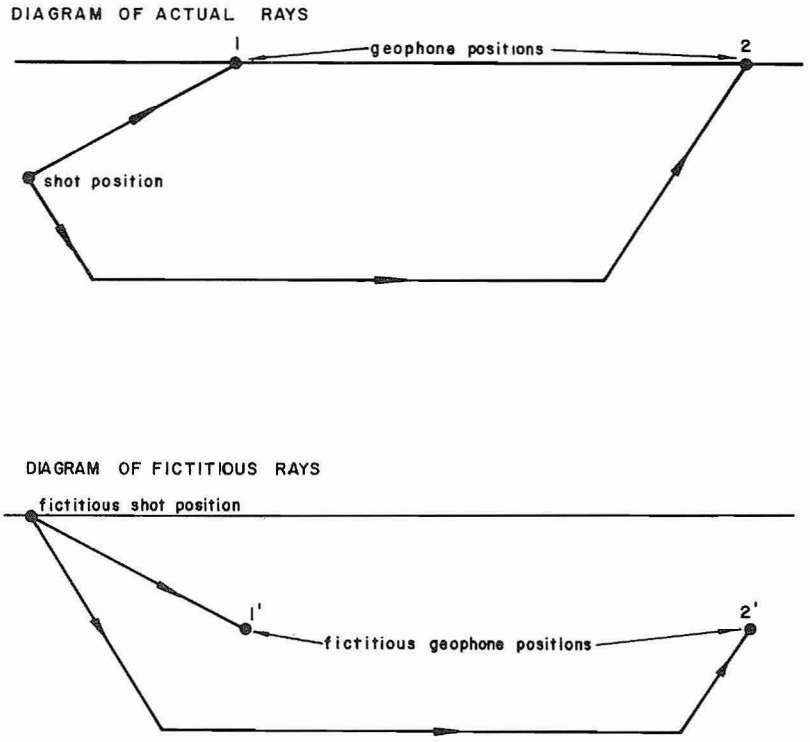


Figure 2. Basis of uphole wavefront method.

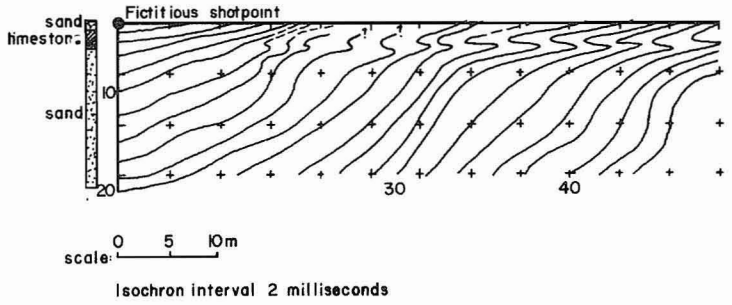


Figure 3. Wavefront diagram for a thin high-velocity layer (after Meissner, 1961).

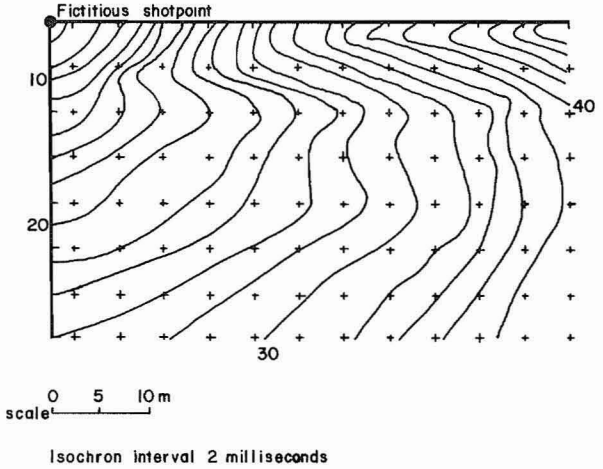


Figure 4. Wavefront diagram for a multi-layer earth (after Meissner, 1961).

difficult, if not impossible. Laboratory measurements on borehole samples or conventional uphole shooting are two other methods which may be adopted.

The usefulness of the uphole shooting procedure may be increased considerably by adopting the method of uphole wavefront investigations as proposed by Meissner (1961). The presence or absence of velocity contrasts may be revealed by this method and, sometimes, the existence of a particular wave type can also be established. In this method, the normal practice of uphole shooting is followed, except that each shot is recorded by a full spread of geophones. The construction of the wavefront diagram is then based upon the times of first arrivals at the geophones. In plotting these times, the position of each geophone is displaced vertically below the surface, to an elevation corresponding to the shot depth, and the position of all of the shots is assumed to be at the top of the borehole (Figure 2). The wavefront diagram is constructed by drawing in the isochrons. This diagram represents completely fictitious wave paths, but as long as velocity is mainly a function of depth, the approximation to the real wavefront diagram is accurate enough for most purposes. When deviations from these velocity conditions exist, they can usually be identified in the diagram (see Figure 5).

The paper by Meissner contains many examples of wavefront diagrams in surficial deposits. All of these diagrams were obtained from plots of first arrival times and three examples are reproduced in Figures 3, 4 and 5. Figure 3 clearly shows a thin, high-velocity layer at a depth of about 2 metres, the presence of which would have been missed in a normal uphole survey. This high-velocity layer correlates with a thin limestone which was encountered in the survey area.

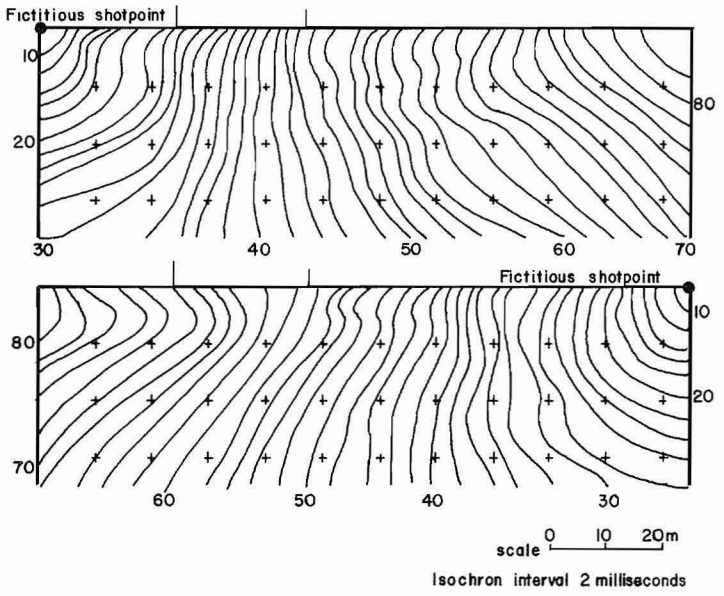


Figure 5. Wavefront diagrams for reversed shots showing near-surface inhomogeneity (after Meissner, 1961).

wavefront diagram is fairly typical of uniform velocity conditions with evidence in the lower part of the diagram, for a headwave coming from a horizon below the depth of investigation.

The wavefront diagram shown in Figure 6 was plotted by the author from results obtained during investigations of glacial deposits in Saskatchewan (Burke, 1968). Below 20 feet, the shape of the wavefronts is characteristic of a direct wave. An abrupt change in the slope of the wavefronts between 50 and 90 feet suggests that a velocity inversion is associated with the bedrock surface at 55 feet. The semiconsolidated till has higher velocity than the underlying bedrock of soft shale. Therefore it is impossible to use the seismic refraction method to map the bedrock profile in this area, a fact which may help to explain some of the previous failures of the method elsewhere in the Prairies (see Lennox and Carlson, 1967). The change in the direction of the slope of the wavefronts above 20 feet is characteristic of a head wave, possibly associated with the silt-till interface.

Sometimes, although there is a lack of contrast in compressional velocity, an adequate contrast exists between shear-wave velocities. The possibility of recording shear or converted waves from such boundaries should therefore not be overlooked. Berzon, *et al.* (1963) have shown in field experiments that it is

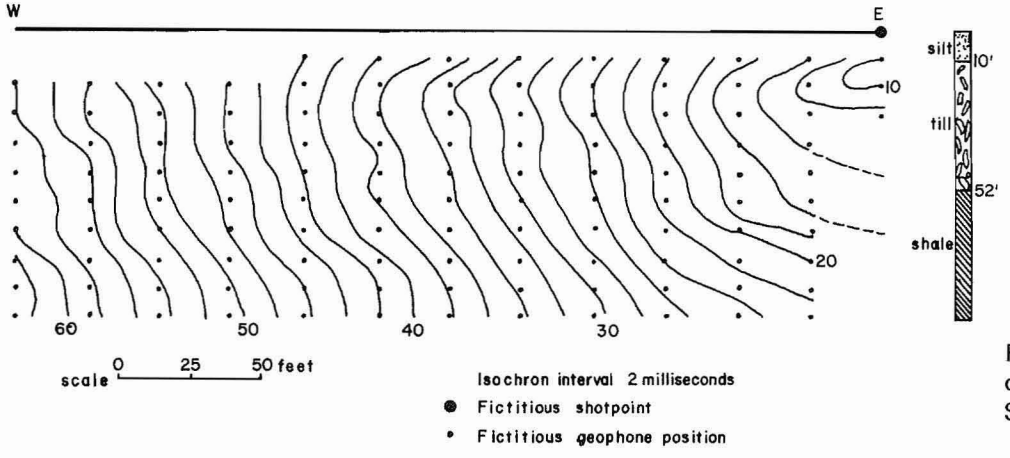


Figure 6. Wavefront diagram for surficial deposits overlying a low velocity bedrock, Stenen, Saskatchewan.

The wavefront diagram for a multilayer situation is shown in Figure 4. A thin, high-velocity layer at the surface overlies a lower-velocity layer between depths of 5 to 20 metres. This lower-velocity layer is underlain by a higher-velocity layer between 20 and 45 metres. A lower velocity below this latter depth is indicated by the bending back of the wavefronts in the lower portion of the diagram. If hammer blows, or shallow shots, had been used to provide the seismic energy, first arrivals from the deeper, high-velocity layer would have been masked by the high-velocity layer at the surface.

The third example (Figure 5) shows wavefront diagrams from shotholes at both ends of a 115-metre spread. The bunching of isochrons, between horizontal distances of 20 and 45 metres in the upper diagram, and their spreading out at the same place, between 70 and 95 metres in the lower diagram, indicates a deviation from the necessary velocity conditions to obtain a true representation of the wavefronts. This deviation does, nevertheless, indicate a near-surface lateral change in velocity which may be of interest in some investigations. The remainder of the possible to record and identify converted reflected waves from boundaries represented by weak velocity contrasts, i.e.,

$$\frac{V_{1comp}}{V_{2comp}} \geqslant 0.7, \quad \frac{V_{1shear}}{V_{2shear}} \geqslant 0.7$$

The best resolution of these waves occurred at distances coinciding with the critical angles of the reflections, where there is a two- and threefold increase in amplitude.

Although points of equal phase are difficult to pick for secondary arrivals on seismograms, Berzon (1964) and Meissner (1965) have extended the use of uphole wavefront diagrams to such events. Berzon shows that the wavefront diagram method can be applied to the study of shear and converted waves and even assist in their identification.

Meissner (1965) identified the existence of shear waves in investigations of near-surface sediments and plotted the appropriate wavefront diagrams. Shear-wave velocities were then used

in conjunction with compressional wave velocities to obtain Poisson's ratio and other elastic constants for the sediments. He also concluded, from this and previous studies, that nearly all near-surface sediments are anisotropic, with 80% of them having a ratio of vertical velocity to horizontal velocity of 1.1 to 1.5.

Amplitude and frequency studies. Development of magnetic recording, and the increase in the dynamic range of recording systems, has made possible the high-fidelity reproduction of seismic waves. This has stimulated interest in the amplitude and frequency characteristics of seismic waves and their potential application in prospecting. Many research groups are now actively engaged in theoretical and experimental investigations of the processes involved in the attenuation of seismic energy. Two recent publications by Howell (1963) and by White (1965) have reviewed much of this work, and the reader is referred to these two publications for a full treatment of this subject.

The causes of changes in the amplitude and pulse shape of seismic waves in its passage through the earth may be conveniently divided into four groups: geometrical divergence of wavefront; absorption; scattering; and diffraction.

The attenuation of amplitude associated with the divergence of a particular kind of wavefront is a function of distance. For example, the amplitude of a harmonic wave, spreading out spherically from a point source, is inversely proportional to the distance travelled. However, for a transient seismic pulse, a wide range of such functions is possible and it is difficult to separate the effects of divergence from those of absorption in field experiments.

Two major absorption mechanisms in rocks have been suggested; solid friction and viscous losses. Solid friction results from the relative movement of grains and crystals in the same rock. Even within the same crystal or grain, defects can also cause frictional losses. Viscous losses occur in porous rock, when the contained fluid moves relative to the rock matrix. White gives a detailed, theoretical discussion of these loss processes. Because rocks differ in their grain and crystal size, porosity and fluid content, a relationship between the absorption coefficient and rock type is expected. Karus (1958) has shown that rocks can sometimes be better differentiated by their absorption coefficients than by their velocities.

Absorption coefficients of different rock types have been measured in the laboratory using ultrasonic pulse and resonance bar methods. Howell (1963) has compiled a table of the values obtained by many workers. Although there is a scarcity of results reported for unconsolidated materials, one interesting result is that obtained by Nyborg, *et al.* (1950) for dry and partially water-saturated soil. They obtained a range of 0.2 to 3.0/cm for the absorption coefficients for dry soils and 3.0 to 6.4/cm for partially saturated soil. This latter range is significantly greater than that reported for more consolidated rocks, e.g., shale 0.79 to 1.8/cm (Krishnamurthi and Balakrishna, 1957). However, values determined in the laboratory should be used cautiously when predicting what happens in the field, because the rock samples may not be representative of the rock *in situ*, and the wave frequencies of the waves are much higher than those produced by explosions in the earth.

White (1965) also includes a review of experimental results related to absorption of seismic energy, but does not report any values for surficial materials. However he does make two comments of particular interest for amplitude studies; attenuation is directly proportional to frequency for travelling waves; and attenuation is increased when moisture is added to an originally dry rock.

Small discontinuities, such as those associated with grain boundaries, will produce a scattering of the main wave types. This conversion of energy is difficult to distinguish from true absorption and has been termed 'pseudo absorption' by Howell. It would be intuitively expected that this process of scattering would cause a large amount of attenuation in an unconsolidated deposit of clastic material. Similarly, diffraction of energy would be an important process if the deposit contained a large amount of angular material. It might be possible to locate these deposits by an anomalous amount of attenuation. The frequency spectrum of the resultant seismic signal might also be related to the range of size of particles present.

O'Brien (1959) has shown there are factors, besides those mentioned, which govern the amplitude and pulse shape of a recorded seismic signal in a field experiment. He classifies these under three headings: changes in the source; effects of geophone to ground coupling; and instrumental distortion.

The results of O'Brien's work suggest that, although these factors have no serious effect on the pulse shape, they can cause much scatter in the amplitudes. Therefore, in any experiment designed to study the dependence of attenuation on the properties of the transmission medium, these three contributing factors must be controlled or eliminated.

In a later paper, O'Brien (1967) has described the systematic procedure which must be followed to use amplitude measurements in seismic refraction surveys. After corrections had been made for the above factors, he found that the residual amplitude was controlled mainly by the nature and thickness of the refractor. Refractors were divided into three groups on the basis of their mean residual attenuation;

(a) Thick non-porous refractor – few hundredths of a decibel per wavelength.

(b) Thick porous refractor – few tenths of a decibel per wavelength.

(c) Thin refractor – few decibels per wavelength.

Although there are many other references relating to the study of attenuation of refracted and direct waves in the literature (see bibliography of O'Brien, 1967), only one reference has been found which makes a study of the absorption properties of a groundwater source in an unconsolidated material. Berson, *et al.* (1959) studied the frequency spectra and absorption coefficients of refracted, compressional waves from a sequence of surficial deposits. The predominent frequency of refracted waves from an aquiferous sand was much higher than the range of predominant frequencies from a clayey layer (200 cps compared to a range of 80 to 100 cps), even though both units have similar velocities. Also, the absorption coefficients of the waves refracted by the aquiferous sand varied over a wide range (0.002 - 0.004 to 0.040/m), while the velocity of the sand had only a range of 1500 to 1800 m/sec. By comparing the observed values of the absorption coefficients with values that were computed for waves in a granular medium, Berson, *et al.* showed that this wide range could be explained by changes in porosity and the size distribution of grains in the sand. Unfortunately they were unable to substantiate the validity of their hypothesis with the reported

experimental results, as no determination had been made of porosity and grain size.

Attempts to confirm the results of Berzon, *et al.* in similar studies of glacial deposits in Saskatchewan have so far been unsuccessful. In a comparison of the frequency spectra of refracted waves associated with a till deposit and a coarse sand and gravel deposit, it was found that major variations were associated more with the shothole conditions than with transmission through the deposits. Although the same charge and shooting depth was used in each shothole, changes in the material at the shot depth could not be avoided due to the heterogeneity of near-surface deposits. Molotava (1964) has shown that lithological conditions at the shotpoint exert a considerable influence on the frequency spectrum of the resultant seismic signal. Therefore it is imperative that uniform conditions exist before studies based on frequency characteristics and absorption coefficients be applied.

The electroseismic effect (seismoelectrical effect of the second kind). The electroseismic effect (seismoelectrical effect of the second kind) is an electrical potential which occurs at the surface of the ground before the arrival of any seismic waves from an explosion (Martner and Sparks, 1959). This effect is believed to be associated with the charged bounded layer of electrolyte moving relative to the oppositely charged pore surface, during the passage of a seismic wave. Martner and Sparks suggested that an electroseismic pulse was generated at some depth below the surface and transmitted from there as an electromagnetic wave. Its time of arrival at the surface should therefore be essentially the same as its time of generation at depth. In the reported field experiments, the generation of the electroseismic effect was associated with the base of the 'seismic' weathered layer and its use in the determination of the depth of this layer was stressed. In a discussion of Martner's and Sparks' paper, Kindij (1960) suggested that the amplitude of the electroseismic pulse might be related to the permeability of the medium, as the latter parameter will control the movement of the contained fluid.

Measurements by Chien-San and Ziangirov (1965) at ultrasonic frequencies of excitation showed that the electroseismic potential has a maximum value of between 1 and 2.4 millivolts, in various types of soils. The maximum value is reached when sufficient electrolyte is present to just compensate the surface charge of the solid framework.

The existence of the electroseismic effect has been confirmed in field tests in Saskatchewan. Preliminary results from detonations of explosives at various depths in glacial till show that the amplitude of the electroseismic signal first increases with depth, and then decreases. This variation in amplitude may be associated with changes in saturation of the till layer, but more tests are needed before such a relationship can be considered as established.

Limitations introduced by small-scale investigations

As the scale of a seismic investigation is reduced, the difference between the theoretical assumptions on which the method is based and the actual physical conditions of the field experiments may become so pronounced, that unreliable and meaningless results are obtained. Because many of the more permeable zones occur in the top few hundred feet, this is a common problem in seismic prospecting for groundwater in surficial deposits. Domzalski (1956) considers most of the important problems associated with shallow seismic refraction investigations. Evison (1952) has also reviewed the limitations of the conventional seismic techniques applied to shallow problems. A brief summary of those factors which are significant in the mapping of surficial deposits will be considered as follows, while the reader is referred to the papers mentioned for a more comprehensive treatment of the whole problem of shallow seismic investigations.

Surveying and planting geophones. Because of the small distances involved, special care should be taken in the setting out and surveying of the spread of geophones for shallow surveys. Rapid variations in seismic velocity of the near-surface material make it advisable to avoid abrupt changes in elevation. Series levelling is usually necessary to provide the required degree of accuracy in the determination of the relative elevations of the geophone positions.

Variation in the composition of the near-surface material, in which the geophones are planted, is also a potential source of error. In a test, Domzalski (1956) obtained a range of approximately 1200 feet/sec in the velocity of a typical group of surface materials. Errors of several milliseconds might therefore be introduced because of changes in 'planting' conditions. In addition, variations in ground-geophone coupling introduce distortion of recorded seismic pulses, especially at high frequencies (O'Brien, 1965). The particle-velocity plots, from three-component geophones, may also be distorted by variations in the ground-geophone coupling between vertical, radial and transverse motions (Bycroft, 1957).

Conditions at the shotpoint. The problems involved in generating distinct bodily waves in unconsolidated deposits have already been discussed. Inhomogeneities at, or very near to, the shotpoint can also cause variations in the travel times of waves. Repeated use of the same shothole will introduce similar problems. Use of different shotholes at the same site can be just as disadvantageous, because inhomogeneities are introduced by previous explosions in adjacent shotholes. Any of these factors, with variations in the size and weight of the explosive charge, changes in the quality of tamping and differences in the rate of detonation, can lead to serious errors.

The resolving power of the seismic method in small-scale surveys is also limited by the use of conventional explosive and hammer sources. These sources generate pulses containing a high proportion of low-frequency energy with a resultant increase in the duration of events on the record. Evison (1952) has shown that the duration of a shallow reflection event generated by a conventional explosive source may cause problems in recognizing later arrivals from a considerable depth below a shallow reflector. He suggested that the duration of the pulse might be shortened by the use of a controlled source of energy. In a later paper (1956), Evison describes an electromagnetic vibrator for this purpose. Pulses of variable frequency and duration were generated and the resulting seismic waves recorded along a 200-foot transverse. Several events were identified as arising from subsurface boundaries, such as a clay-siltstone interface. Surprisingly, reports of the use of such a source for shallow seismic surveys are not more numerous.

Near-surface inhomogeneities. The uppermost layer of surficial deposits is commonly composed of a loose, unconsolidated material with a low velocity. Although the thickness of such a layer varies by only a few feet, the relatively low velocity can introduce significant errors into the timing of events. For example, a layer with a velocity of 1000 feet/sec overlying one of 5000 feet/sec will introduce a time difference of 8 milliseconds for an increase of thickness of 10 feet in the first layer. It is usually impossible to predict, or impracticable to measure variations in thickness of this order of magnitude. An error of 8 milliseconds assumes detrimental proportions, if the total time of travel is only some tens of milliseconds. Rugged topography aggravates this problem, since low-velocity deposits tend to occur in depressions and are absent on high ground. The use of uphole wavefront diagrams can sometimes assist in the identification of near-surface inhomogeneities, as the previously mentioned example of Meissner demonstrates (Figure 5).

Conclusions

1. Much of the energy from hammer and explosive sources in unconsolidated materials is transformed into various types of surface and coupled waves. Incorrect depth estimations and erroneous identification of deposits may result if any of these surface or coupled waves are mistaken for low-velocity compressional waves. Therefore the identity of an event must be established on a seismic record, before making an assumption that it is a refracted or reflected event, and before it can be used in a standard interpretation procedure. To aid in this identification, it is suggested that three-component recordings and uphole wavefront investigations be made to determine the mechanism of propagation of a particular event.

2. The seismic velocity in a surficial deposit is a complex function of many factors; the velocity in lithologically identical deposits varies with depth of burial and degree of saturation. Therefore a study should be made of the *in situ* velocity model with an uphole wavefront investigation. The resultant wavefront diagram often reveals the presence of thin, high-velocity or low-velocity layers or a velocity inversion. These latter features would be difficult to isolate using the more conventional methods of investigation.

3. The abrupt increase in compressional wave velocity as a medium becomes fully saturated, has been used as the basis for successful determination of the water-table. However the success of this method will depend on how closely the boundary between the partially saturated and fully saturated zone approximates the position of the water-table. Under conditions of low overburden pressure, the role of moisture suction also may become significant and the compressional velocity increase instead with the desaturation of the medium.

4. The use of amplitude and frequency characteristics in seismic investigations requires ideal experimental conditions. Berson, *et al.* (1959) have successfully used this procedure to distinguish an aquiferous sand from a clayey layer with the same velocity. Their results also suggested a relationship between the porosity and the absorption coefficients of arrivals from the aquiferous sand. Unsuccessful attempts were made to establish similar relationships for a sand and gravel aquifer in Saskatchewan. The results obtained were erratic, with the amplitude and frequency spectrum of the signal being strongly influenced by slight changes in the shotpoint conditions. A controllable source of energy is recommended in future studies.

5. Because of its possible relationship with the permeability of materials, the electroseismic effect should be more thoroughly studied.

6. The relatively small scale of many seismic investigations in surficial deposits considerably magnifies the errors introduced by inaccurate surveying, inadequate precautions in layout and poor operational procedures. Extra care should therefore be taken to minimize or eliminate the cause of such errors.

Acknowledgments

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Discussion on A review of some problems of seismic prospecting for groundwater in surficial deposits, by K.B.S. Burke

Hall, D.H., University of Manitoba, Canada. Perhaps one way to cut down the seismic cost is to go to a blanketing arrangement, time-term approach, instead of straight lines which is really very inefficient. I understand the University of Western Ontario has prepared a method of time-term approach in near-surface problems, i.e., instead of having a straight profile, they use a central shot point and record out in different directions. What are your ideas on that?

Burke, K.B.S. In the Prairies it is difficult to get away from the 90° grid roads. However, I do not know what the results were of this work.

Hall. It would be interesting to investigate the economics of the fan. Is anyone here from the University of Western Ontario?

Utard, M., Geoterrex CGG, France. It is a beautiful technique but you will need to know all parameters. If you wish to shift from one marker to another one you miss the complicated structure. But I agree, if you plot very carefully seismic soundings, you could relate it to arc shooting.

Hall. The new name 'time term' goes back to the early days of refraction of arc shooting.

Gendzwill, D., Saskatchewan Research Council, Canada. Would it be fair to say that the conclusion of your paper is that seismic resolutions of sand and gravel aquifers in the western Prairies is very poor?

Burke. Using the conventional techniques, I tend to agree with you. Something that we have not tried, though, is to use a controlled source of energy.

Hall. One thing that impressed me in your paper was that, looking from the point of view of seismology, in general, the near-surface problem seems to be inspiring people to develop new techniques, new kinds of waves that you were speaking of, etc. Do you think that these new techniques will develop into a new body of information and can you predict where this excitement might be?

Burke. Of course, wave-front diagrams have been used in Europe for 10 to 15 years now. Electroseismic effects have also been known for a long time. Hopefully, they will provide information that is new in the area of surficial prospecting but it is difficult to predict exactly where these advancements will be.

Hall. I have the feeling that although these investigations have been used before, they are being brought into new focus, at least in Canada, and I think that this is a very interesting development in our country.

Collett, L.S., Geological Survey of Canada. I think that this is what is happening in the borehole logging field. Logging has been used in the oil exploration industry for a number of years, the near-surface geology has been neglected and now we are just beginning to apply these techniques to groundwater problems.

Wyder, J.E., Dept. Energy, Mines and Resources, Canada. Dr. Burke, in Don Lennox's paper there seems to be some problem of correlating the hammer seismic with conventional seismic results, particularly in the near surface. Would you care to comment on this?

Burke. If we were to plot the particle motion hodographs for some of the hammer seismic waves, the way we do waves from explosive charges, I think you would find that the hammer seismic waves are not really compressional waves. The hammer blows generate some type of surface wave. Since they are not the same kind of waves, we would not then expect the velocities to be the same.

Wyder. Have you looked into the relationship between the depths calculated using hammer seismic and those depths using conventional seismic techniques? In other words, although one is using the wrong theory, does one come up with the right answer?

Burke. I would not know without looking into the matter more fully. Obviously they get the right answer from hammer surveys in many circumstances. The thickness of near-surface material is often so small that the error introduced may be within the experimental error of measurement, anyway.

Interpretation of geoelectrical resistivity measurements for solving hydrogeological problems

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Abstract. Many endeavours have been made to fully utilize geoelectrical sounding graphs recorded by resistivity (four-electrode array after Schlumberger or Wenner) in the interpretation. It is known that 2- and 3-layer diagrams that are published in the form of sets of master curves allow, in combination with auxiliary point diagrams, an approximate physical interpretation provided the principle of equivalence is sufficiently observed for the geological situation, and that good field measurements which are not influenced by secondary effects are obtained. Moreover there is, nowadays, the possibility for a rapid and exact computation of master curves for horizontal *n*-layer cases (*n* unlimited) with arbitrary resistivities. This permits a full utilization of the total information content of a measured graph. Case histories from humid and arid areas illustrate how, by geoelectrical resistivity measurements, hydrogeological problems may be cleared up using a suitable combination of the presently available methods of interpretation. This refers, in particular, to the search for and the delimitation of aquifers and the fixation of the freshwater/saltwater boundary in the underground.

Résumé. Au cours des dernières années on a maintes fois essayé d'utiliser des enregistrements graphiques des sondages électriques (disposition des électrodes en quatre points d'après Schlumberger ou Wenner) pour les faire servir dans leur ensemble à l'interprétation. Nous savons que les diagrammes à deux ou trois couches qui sont publiés sous forme de jeux d'abaques permettent, en combinaison avec des diagrammes de points auxiliaires, une interprétation physique approximative à condition que le principe de l'équivalence soit suffisamment respecté quant à la situation géologique et qu'il existe des mesures nettes non faussées par des effets secondaires. De plus, il est maintenant possible de calculer de façon rapide et exacte les abaques pour les cas à *n*-couches horizontales (*n* sans limite) et comportant des résistivités arbitraires. Cela permet d'utiliser pleinement les renseignements enregistrés sur un graphique. Au moyen de cas précis comportant des zones humides et arides, l'auteur indique comment, par les mesures de la résistivité géoélectrique, les problèmes d'ordre hydrogéologique peuvent être réglés à l'aide d'une combinaison convenable des méthodes d'interprétation actuellement disponibles. Cela se rapporte en particulier à la recherche et à la délimitation des nappes aquifères et des eaux douces et salées souterraines.

The interpretation of geoelectrical resistivity measurements for hydrogeological problems is substantially determined by the physical parameter of specific electric resistivity ρ (Ωm) of the rock. In the treatment of hydrogeological problems we have to deal exclusively with the ionic conductivity of the soil and this, again, depends on the pore or interstitial volume of the rock, on its water saturation and on the electrolytic content of the groundwater. This, as a matter of fact, applies only to rocks in which the matrix can practically be regarded as an insulator. As is well known, clays and marls are an exception which in many cases greatly aggravates the interpretation and requires drilling activities.

A simple general view of the rock resistivities as they occur in the treatment of hydrogeological problems is given in Figure 1. Represented on a logarithmic scale are the resistivities of water, of aquifers, and of impermeable beds which either form the substratum or are intercalated as separating horizons. For pure sands and gravels, the formula by Archie on the upper right shows the connection between water and rock resistivities. The grain size does not occur in this formula. Thus, the geophysicist is unable to give reliable statements on the permeability of an aquifer, that being an essential function of the respective grain size. We know of many cases where very clean water-filled sand bodies were determined by geoelectrical measurements but,

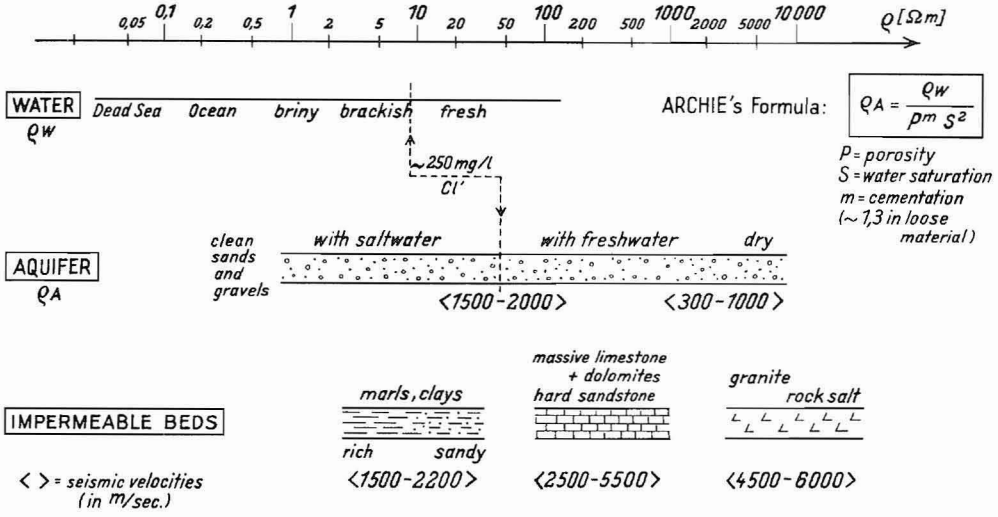


Figure 1. Electric resistivities and seismic velocities.

owing to their fine-grain quality, could only to a limited extent be considered as a water-producing horizon.

One look on the resistivity scale shows that fresh-water-bearing sands and gravels possess medium resistivities. An aquifer of such composition, therefore, can easily be delimited electrically from clayey-marly impermeable beds or from a compact bedrock, but not so, however, from lime- and sandstones because their resistivities have the same order of magnitude.

The dependence of the specific resistivity on the electrolytic content of the pore water renders the geoelectrical resistivity method an effective means for fixing the limit between fresh and salty groundwater. In this connection, important results have been achieved in the past, all the more so, because drilling alone can determine this limit. The problem for the interpretation lies in the good conductivity of the clays and marls which from electrical measurements cannot be distinguished from salt-water-bearing sands and gravels unless the decision can be made on the basis of the general hydrogeological picture. Frequently a boring is unavoidable but generally one boring is sufficient if located at a suitable place. Some case histories on this subject given in 'Case Histories' will underline these remarks.

In addition, the scale shows a resistivity variation between dry and water-saturated loose sediments. The problem of the groundwater level causes a lot of difficulties for the physicist. For a loamy covering very often it cannot be solved without drilling. This, too, shall be shown in the case histories.

The method of seismic measurements can contribute considerably towards a decision in electrically equivalent cases. Since it is now possible in seismic activities to manage without the use of explosives to obtain information on the depth to several 100 m by weight dropping and a registration with a photographic recording system, we are offered a combination of two different methods. Both permit the analysis of a horizontally stratified underground. One look at the seismic velocities in Figure 1 makes evident at once that in cases of equal specific electric resistivities the velocities are strikingly different. This combination of the two methods is the up-to-date domain for geophysicists in solving hydrogeological problems. Through the possibilities offered by electronics we shall soon be in the position to have both these methods at our disposal by mounting the equipment in only one cross-country field vehicle.

As is well known, the interpretation of geophysical measurements is done in two steps: first, from surface measurements, the physical properties of the underground can be derived by a mathematical-physical interpretation; and secondly, this interpretation is correlated to the geological, here hydrogeological, structure. The result, then, is the working material for the geologist, in the form of vertical and horizontal sections, or in diagrams showing the depth location of interfaces between permeable and impermeable layers, salt water and fresh water, etc.

The significance of the geoelectrical resistivity method, as already mentioned, is that it is capable, side-by-side with the seismic method, to subdivide a horizontally stratified underground. Moreover, for a horizontally stratified underground where the resistivity ρ is solely a function of the depth z, there exists a unique connection $\rho_a(L/2) \leftrightarrow \rho(z)$, viz., between the true resistivity and the apparent resistivity which is dependent on the electrode distance L and which is measured at the surface using either the Wenner or Schlumberger electrode configuration.

In 1933, Slichter referred to this and Buchheim (1947) has described a method for the direct computation of $\rho\,(z)$ from $\rho_a(L/2)$ by continuous integration processes. However, in practice one still has to rely, now as ever, on the application of theoretically computed master curves, for which Stefanescu, *et al.* (1930) have given the integral presentation

$$\rho_a\,(L/2) = \rho_1(1 + 2\,(L/2)^2 \int_0^\infty \theta(\lambda) J_1\,(\lambda L/2)\,\lambda d\lambda)$$

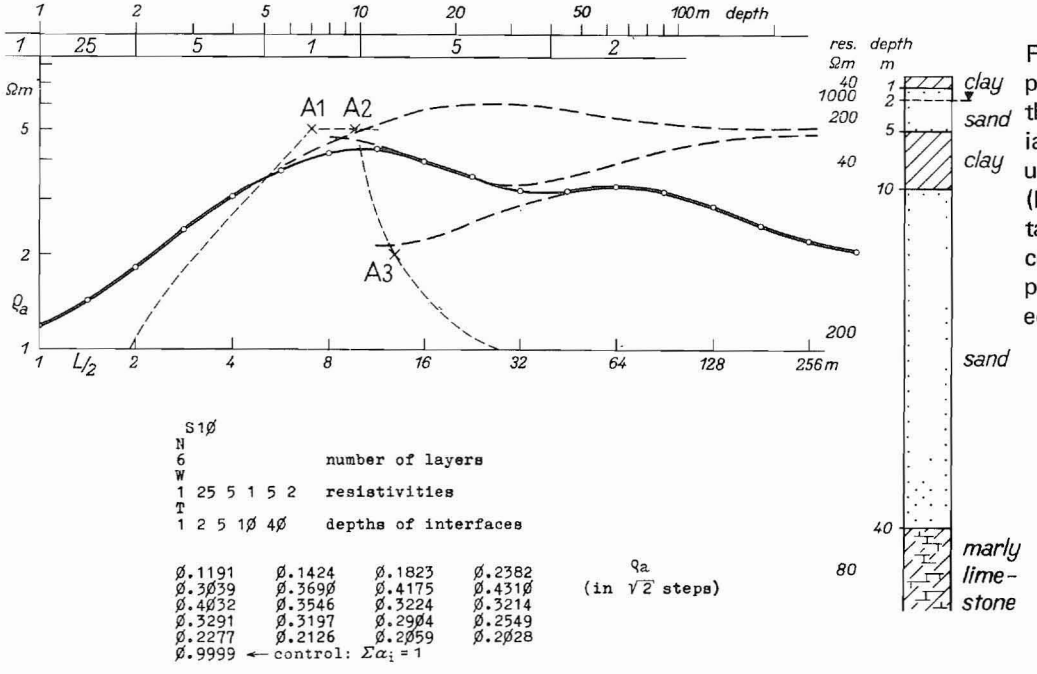


Figure 2. Six-layer master curves computed after Deppermann (1968). Besides the accurately computed curve, the auxiliary points A 1-3 and the 3-layer graphs used for interpretation are plotted (Homilius, 1961). The $L/2$ scale is simultaneously used as depth scale. In the left corner, the result of the electronic computation is represented as originally printed by the computer shown in Figure 4.

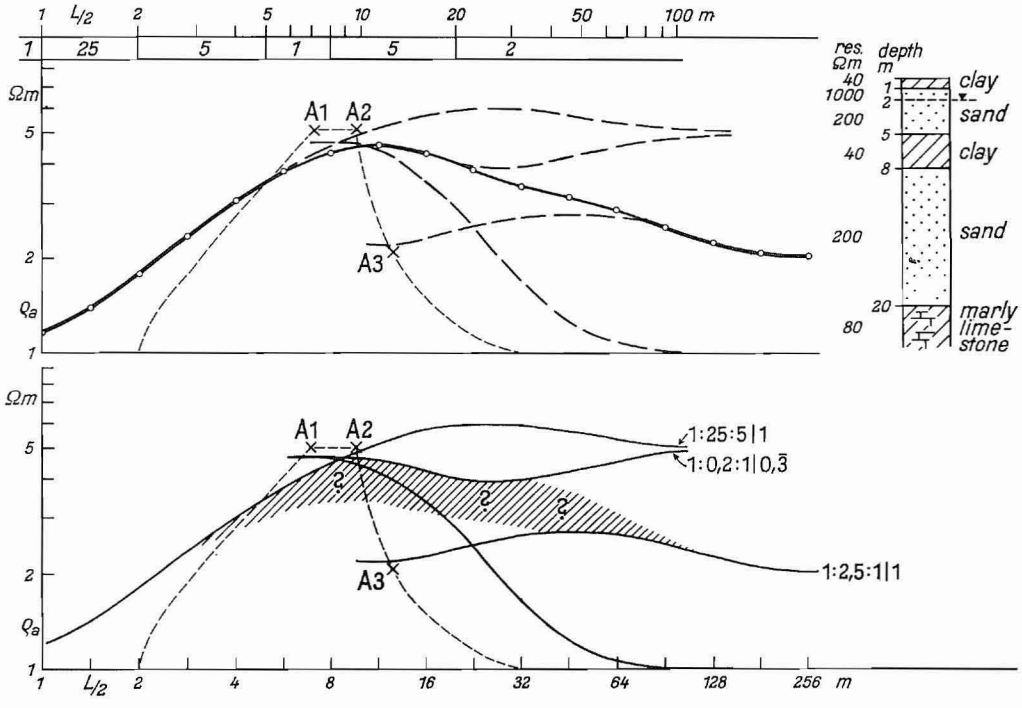


Figure 3. This 6-layer master curve should be compared with Figure 2. Here the second aquifer storey has a thickness of only 12 m. This means that by using the auxiliary point method only part of the whole information content is covered by the 3-layer graphs used for interpretation (see upper section). In the lower section, only the auxiliary points and 3-layer graphs are plotted to demonstrate the difficulty for constructing the exact 6-layer graph by the auxiliary point method. (1:0.2:1/0.$\bar{3}$) means that a 3-layer curve is used with the resistivity ratio of 1:0.2:1 for the three layers, and a thickness of 0.$\bar{3}$ for the second layer compared with the top layer.

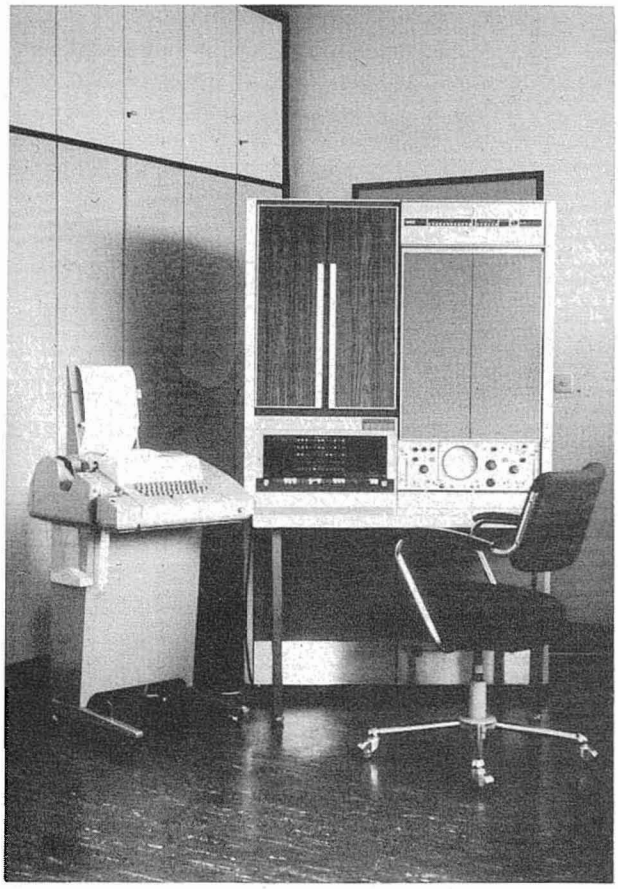


Figure 4. On-line computer for calculating geoelectrical measuring curves (only the left part of this instrument is used for computation).

Since 1955 this integral renders the possibility of interpreting arbitrarily multilayered cases, after a recurrence formula for the kernel function θ had been found (Flathe, 1955 a, b).

Master curves are published in catalogues for 2-, 3- and 4-layer cases (Compagnie Générale de Géophysique, 1955; Orellana and Mooney, 1966). A preliminary analysis of measured curves can be made during field work, whereby the so-called 'auxiliary point methods' are applied. Of course, it must be kept in mind that this kind of interpretation can only utilize part of the measured data. The entire information content of a sounding graph can comprehensively be obtained only by exactly computed model graphs that in their whole course correspond to the measured graph. Figures 2 and 3 may illustrate this fact. Theoretically computed 6-layer graphs are represented together with the auxiliary points and 3-layer graphs used for the interpretation. Details are given underneath the figures. The comparison shows clearly what is omitted if the total graph is not exactly computed.

The computation of these multilayered curves no longer requires a big electronic computing centre. Deppermann and Homilius (1965) have developed a method of calculation which can be applied, practically beside their desks, on an on-line computer and which, for instance for a 6-layer graph, needs less than 1 minute working time. This computer is shown in Figure 4. Since actually only part of this electronic equipment is really used for the computation of the master curves, it is possible to carry this small computer along in the field vehicle. Everything necessary is available during field work: equipment for electric and seismic measurements and the electronic computer. This is an immense advantage for a team working far away from a computing centre.

Nevertheless, this mathematical-physical capability constitutes only the first step on the way: the exact conformity of measured graph and master curve does not yet present the evidence for the final hydrogeological situation. This hydrogeological interpretation is the decisive problem confronting the geophysicist

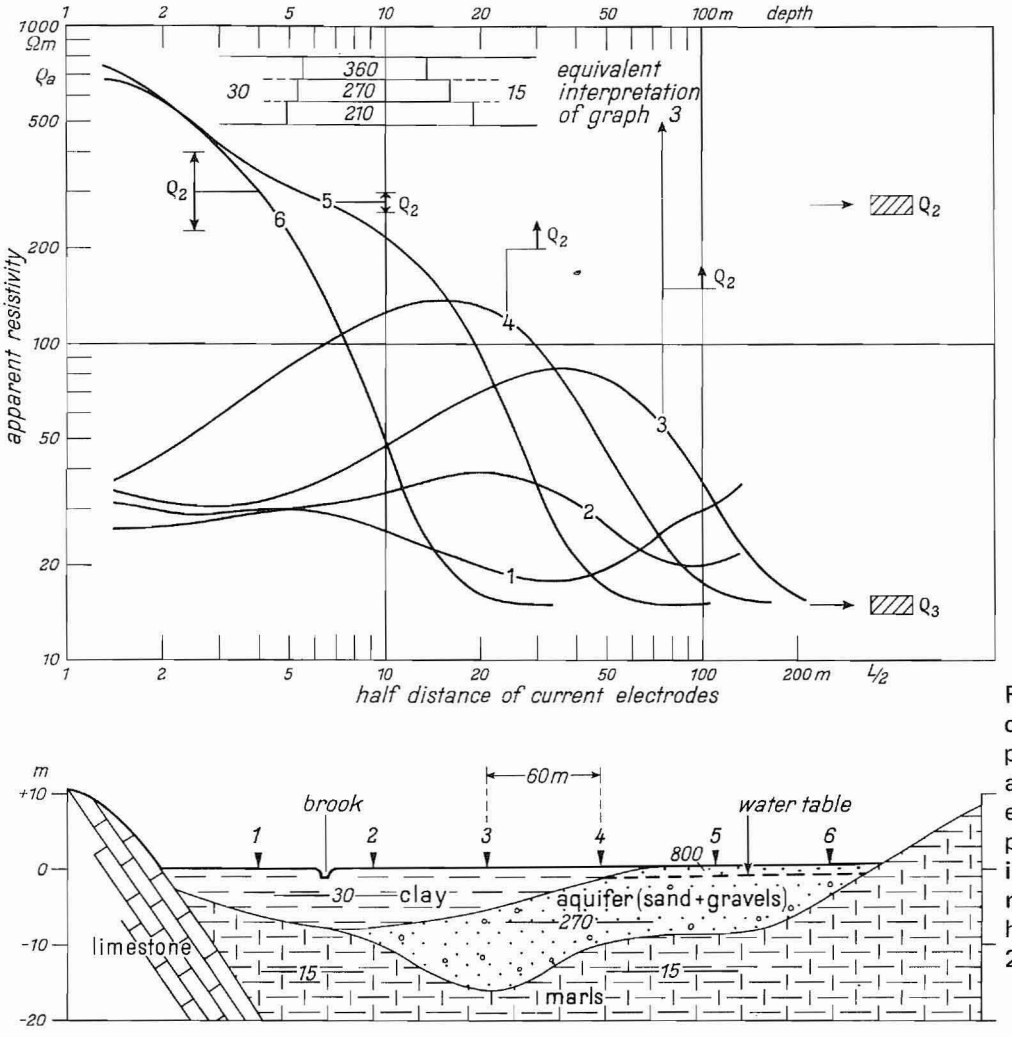


Figure 5. Valley with aquifer. This vertical section shows the result of the interpretation of six sounding graphs plotted above. In this special case the principle of equivalence leaves only a small space for possibilities in the interpretation as is indicated for the values ρ_2 and ρ_3. This result should be compared with the case history of the Upper Kabul Basin (Figure 24).

and which can be solved only in closest cooperation with the geologist. Without a hydrogeological concept and without the complete and absolute utilization of all available hydrogeologic data and records, such as drilling and prospecting logs, water analyses, measurements of the water level, etc., the problem can seldom be solved satisfactorily. One must refer again to the resistivity scale in Figure 1.

Let us study the procedure of interpretation

$$\rho_a(L/2) \rightarrow \rho(z) \rightarrow \textit{hydrogeological facts}$$

by means of a simple example demonstrated in Figure 5. A valley where an aquifer is assumed to exist shall now be delimited. Six soundings were carried out, with the electrode spacing parallel to the valley axis in the direction of the strike. On the left part of the valley, limestones are exposed, while opposite marls are outcropping. It could be expected that the substratum is of marly nature. The measured graphs show on the surface of the left part of the valley low resistivities (<30 Ω m) that are obviously due to a loamy overburden. The high resistivities of about 800 ohmm measured on the right part of the valley indicate dry gravels. How can we get an idea about the shape of the aquifer? The fact is that this measured profile can be regarded as an almost ideal case because on one part the measured graphs show high and on the other part low resistivities for the overburden. A comparison of all the six curves results in a narrowing down of the resistivity for the aquifer to a range a little below 300 ohmm. The tolerance according to the principle of equivalence is given in graphs 3 to 6. The evaluation is based on 270 ohmm for the aquifer. In graphs 1 and 2, the limestone becomes noticeable through a lateral effect (ascent of the rear curve branch).

The ideal case illustrated here was meant to demonstrate how several measurements may furnish an unobjectionable result. No drilling activity is necessary to confirm it. Curve 3 alone would quantitatively not have yielded a satisfactory result as is proved by the indicated equivalent possibilities of interpretation.

However, this case is an exception. The conditions become complicated if the resistivity of the aquifer changes because of a change in the water-chemism or if the groundwater level lies at an unsuitable depth. Both these cases shall be discussed briefly. The two curves represented in Figure 6 were measured in northwest Germany in a dune area. The rather high resistivities of the upper graph showed that the groundwater was a poor electrolyte, after it was evident from the hydrogeological situation that the sands should be saturated with water. The lower graph shows a clear minimum. Initially it was assumed that the cause was a clay

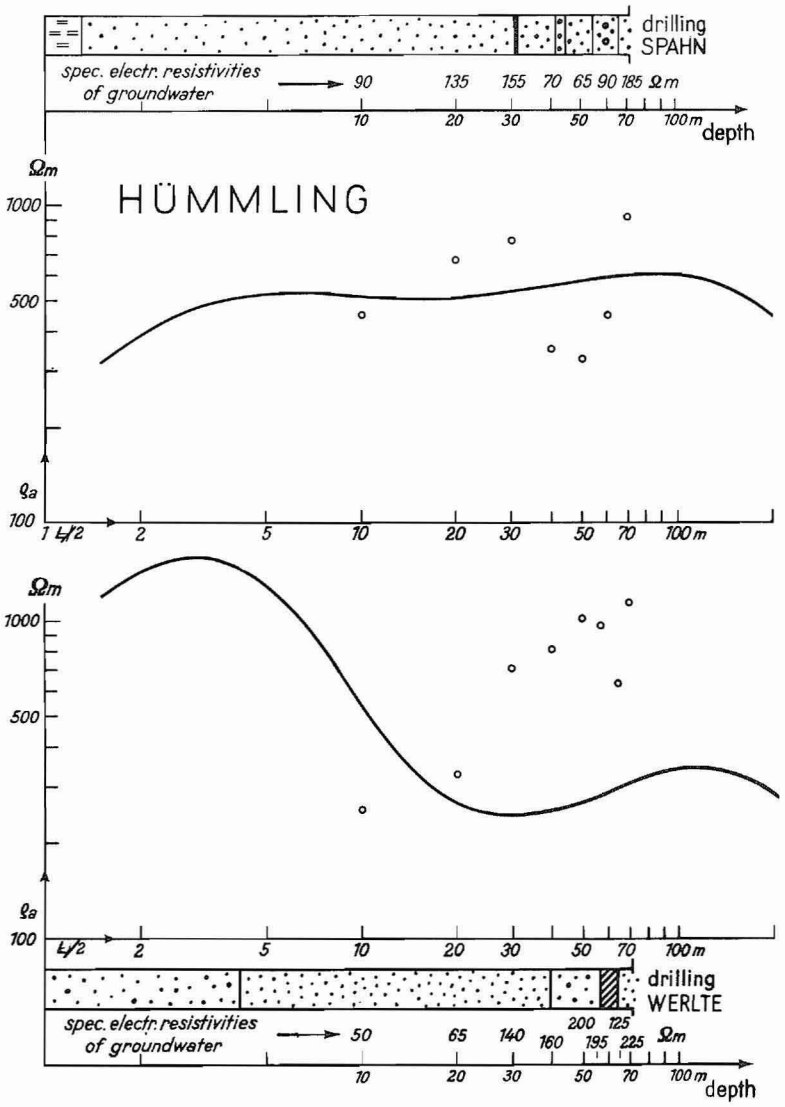


Figure 6. Water chemism influences the specific resistivity. Both measuring curves are based on an equal petrography. The difference in resistivity depends only on the water chemism: the recorded water resistivities, multiplied by a factor 5 (from the Archie-formula in Figure 1) are plotted in the diagram (O).

component which decreased the permeability. But two exploratory boreholes then proved that only the chemism of the water had effected the differences in resistivity. A comparison of the measured apparent resistivities and the resistivities of the groundwater, pumped every 10 m during the drilling process, confirms this fact. Details are given in the figure caption.

The problem of the groundwater level lies in the uncalculable change in the resistivity of dry sands and gravels above the groundwater level. If, additionally, there is a well conducting overburden, then the measured graphs representing a 4-layer case (loamy overburden – dry gravels – aquifer – substratum) have the appearance of 3-layer graphs. Deppermann and Homilius (1965) have studied this subject.

Even if there is sufficient information about the depth of the groundwater level and of the resistivity of the aquifer (from measurements of the water resistivity and using the Archie formula), with an electrically good conducting overburden difficulties were met when trying to determine the depth of the substratum. The graphs show the habit of 3-layer curves although actually 4 layers are concerned. Dry and water-saturated sands and gravels appear as one layer, with a pseudo thickness m* and the pseudo resistivity ρ^*

$$\rho^* m^* = \rho_2 m_2 + \rho_3 m_3$$

Thereby ρ_2 and m_2 denote resistivity and thickness of the dry gravels, ρ_3 and m_3 resistivity and thickness of the aquifer. Determinable is $\rho^* m^*$. Variations in ρ_2 aggravate the determination of m_3 even if ρ_3 and m_2 are known. Figures 7 and 8 show equivalent confrontations of such cases demonstrated in drawn master curves and computed values. Details are given with the figures.

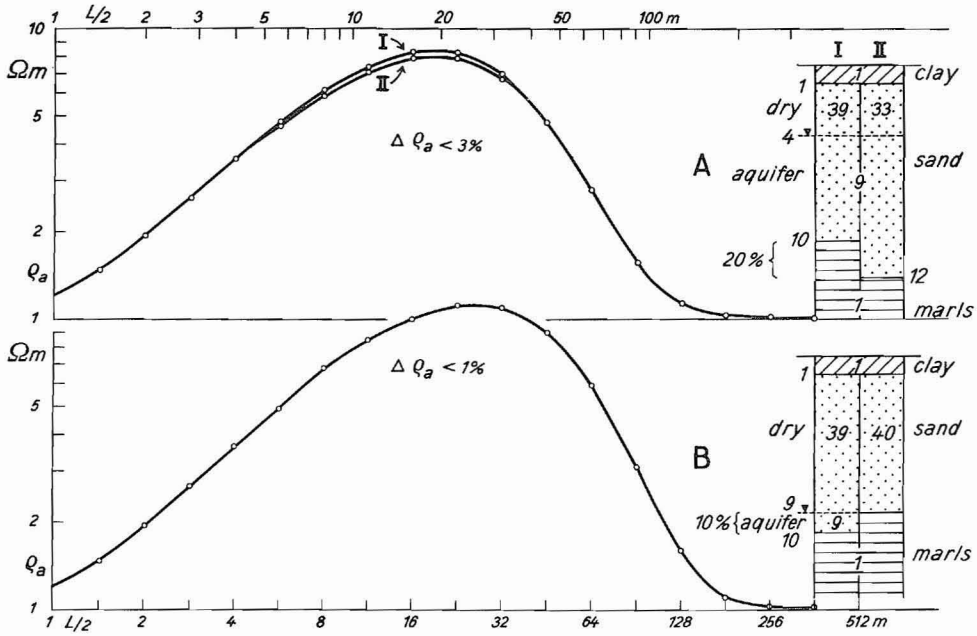


Figure 7. Equivalent graphs corresponding to the groundwater level. A. Influence of the difference in resistivity in the dry sands above the aquifer on the depth of the deeper-seated surface of the impermeable substratum. B. Equivalence of a 3- and 4-layer case. An aquifer below dry sands cannot be electrically determined.

S1		S2		S3		S4
N 4		N 4	number of layers	N 4		N 3
W 1 39 9 1		W 1 33 9 1	resistivities	W 1 39 9 1		W 1 40 1
T 1 4 10		T 1 4 12	depths of interfaces	T 1 9 10		T 1 9
0,1208		0,1205	ϱa	0,1209		0,1210
0,1467		0,1461	(in √2-steps)	0,1471		0,1472
0,1926		0,1913		0,1938		0,1939
0,2617		0,2592		0,2649		0,2653
0,3546		0,3499		0,3631		0,3637
0,4707		0,4625		0,4921		0,4931
0,6057		0,5897	Δϱa = 0.0067 (< 1%)	0,6882	<	0,6949
0,7286		0,7096		0,8370		0,8402
0,8136	>	0,7925	Δϱa = 0.0211 (< 3%)	0,1001		0,1003
0,8102		0,7925		0,1109		0,1110
0,6875		0,6794		0,1087	=	0,1087
0,4765		0,4785		0,8973		0,8954
0,2713		0,2780		0,5918	>	0,5892
0,1532		0,1575		0,3101		0,3085
0,1122		0,1135		0,1592		0,1584
0,1041		0,1045		0,1118		0,1116
0,1019		0,1020		0,1037		0,1037
0,1009		0,1010		0,1017		0,1017
0,1004		0,1004		0,1008		0,1008
0,1002		0,1002		0,1004		0,1004

Figure 8. The table gives the exact values calculated by the computer and plotted in Figure 7.

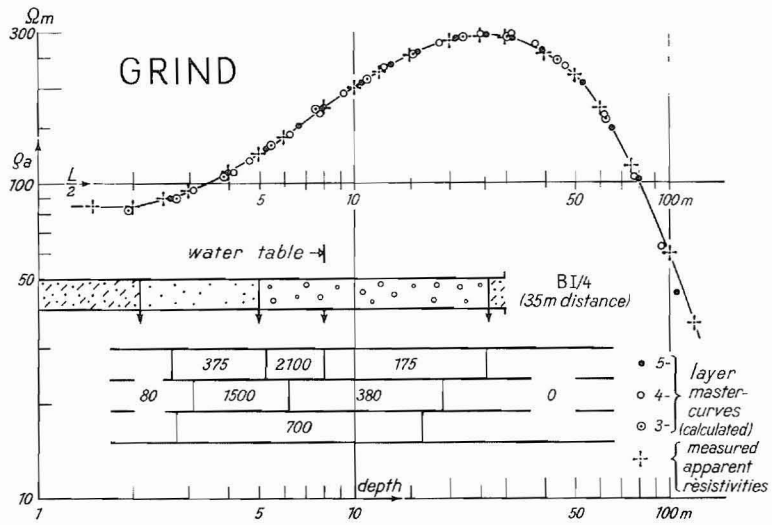


Figure 9. Equivalence between a 5-, 4- and 3-layer interpretation of a sounding graph in the Grind near Düsseldorf/Lower Rhine Valley, Germany (the $L/2$ scale is simultaneously used as depth scale).

The problem of the groundwater level and the determination of the number of layers very often considerably complicates the interpretation of geoelectrical measuring graphs. The sounding graphs are smooth. For this reason we cannot, as with seismic travel time curves, directly recognize the number of layers, and grave mistakes may creep in during interpretation. This may be demonstrated by Figure 9. In a water-development terrain, completely explored by boreholes, which is situated near Düsseldorf at the River Rhine, the depth position of the impermeable substratum was to be determined as exactly as possible. With the application of geoelectrical techniques, this problem could not be solved because the heterogenous structure of the loose sediments prevented the wanted accuracy of statements. The graph was recorded in the proximity of a borehole the profile of which is plotted here. Quite obviously this is a 5-layer case: loamy-fine sandy overburden, thickness 2 to 3 m (80 ohmm), then dry clayey fine sands (some 400 ohmm), dry coarse sands (about 2000 ohmm), and below the groundwater level sands with 175 ohmm, underlain by an almost ideally conducting substratum. The resistivity of the water-saturated sands with 175 ohmm is comparatively low. Without the drilling, the resistivity would have been estimated to amount to 300 to 400 ohmm, merely on the basis of empirical values obtained in

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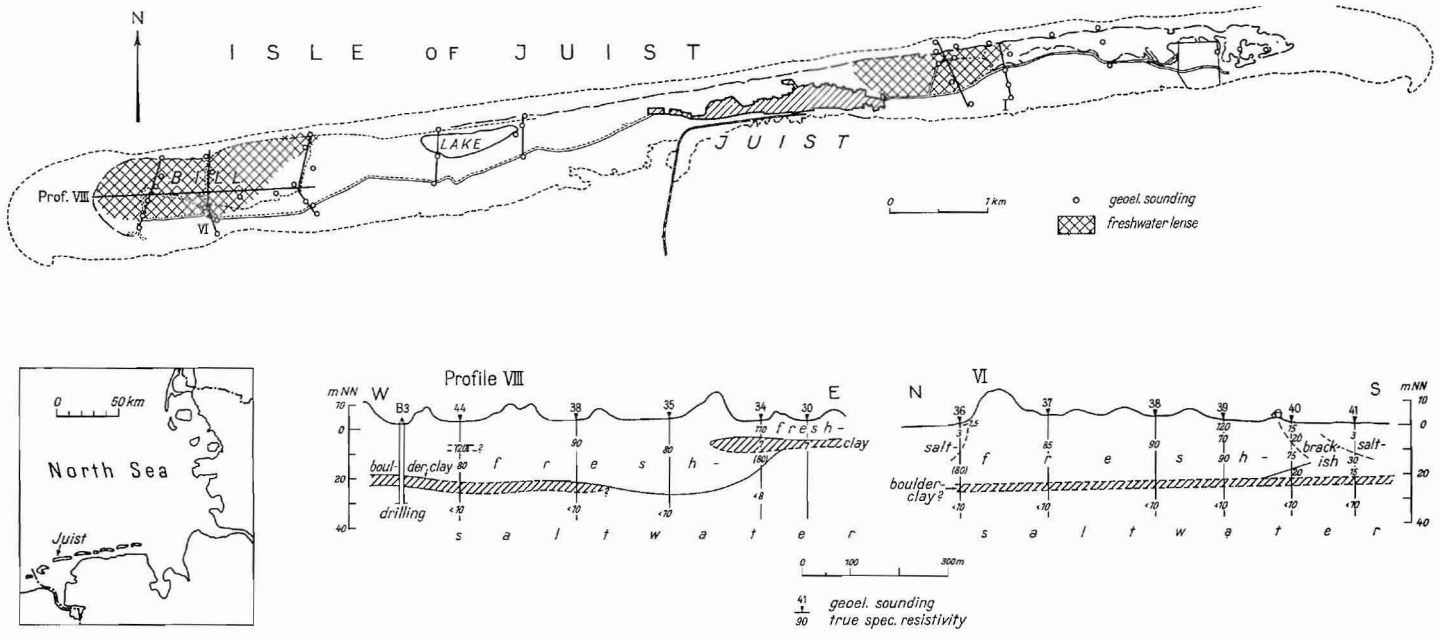


Figure 10. Freshwater lenses on the Isle of Juist.

the prospection area, and the interpretation would have been made on the basis of a 4-layer case: 80 - 1600 - 380 - 0 ohmm. The basement thus shifts upwards for 7 m, and it would further be lifted for 2,5 m if the given curve were interpreted mathematically-physically without any knowledge of the hydrogeology. The graph represents, in fact, a classical 3-layer case. The computed and measured data leave no doubt.

Case histories in the next chapter will demonstrate the application of the discussed principles on practical subjects.

Case histories

Complementary to the fundamental explanations in the previous section, some case histories shall be described with special emphasis on groundwater salinity, in coastal regions as well as inland. For this purpose examples have been chosen from humid, semiarid and arid areas, and we have tried to outline the decisive aspects.* An attempt is made to compare the recorded sounding graphs in the different climatic provinces and continents to find out to what extent there exists a difference or a conformity. All the graphs displayed here have been recorded with the same type of instrument and interpreted by scientists of one and the same institute. Thus it can only be a true comparison and this is the reason why citations from literature have been dispensed with. Naturally, the author may claim to have a good knowledge of this literature, but he sees no useful comparison because many papers communicate results but they withhold the measuring data or give them inadequately. So a real evaluation with equal weight appears impossible. In the case histories an attempt is made to directly compare measuring data and measuring results so that the utilization of hydrogeological knowledge is taken as the most important means for interpretation, using the techniques of mathematical-physical interpretation which were the primary subject of the previous section. The following classification has been chosen for the case histories: (1) Coastal salinity (2 examples); (2) Groundwater salinity inland (2 examples); and (3) Search for and delimitation of an aquifer (1 example). For the third, a multitude of case histories could be drawn here. The principles of such cases, however, are already outlined in the description of a geoelectrical investigation of a river valley in the previous section.

*Within the framework of Technical Assistance for Developing Countries rendered by the Federal Republic of Germany, the Bundesanstalt für Bodenforschung (Geological Survey) Hannover has been performing, already for a decade, geoelectrical investigations for developing useful groundwater. The measuring instruments required for this were constructed and developed in their own laboratories. Even under extreme field conditions, they permit the measurement of voltages as small as 1 microvolt (10^{-6} V).

Coastal salinity.

Freshwater/saltwater interface in loose sediments, Isle of Juist in the North Sea, Germany. The elongated narrow Isle of Juist is situated offshore from the German coast along the North Sea (Figure 10). The substratum consists of Pleistocene sands with good permeability in which the infiltrating rainwaters lead to the formation of freshwater lenses. These freshwater lenses 'float' on top of the specifically heavier salt water which, penetrating from the sea, fills the sands in the substratum of the isle. The extension of these freshwater lenses depends on the quantity of precipitation, the permeability of the sands, and the distribution of impermeable beds in the underground.

For a homogeneous sand body, the fresh/saltwater interface is primarily a function of the difference in the specific weight of the waters, secondarily of the local flow conditions. A classic example is reported by Deppermann, *et al.* (1961); it is the study of the freshwater lens on the neighbouring Isle of Norderney. Conditions on the Isle of Juist, however, are more complex, since in the substratum there occur near the surface clays preventing an infiltration (for example, west of the village) as well as a boulder clay of a few meters' thickness at depths of 20 to 25 m which was drilled at many places.

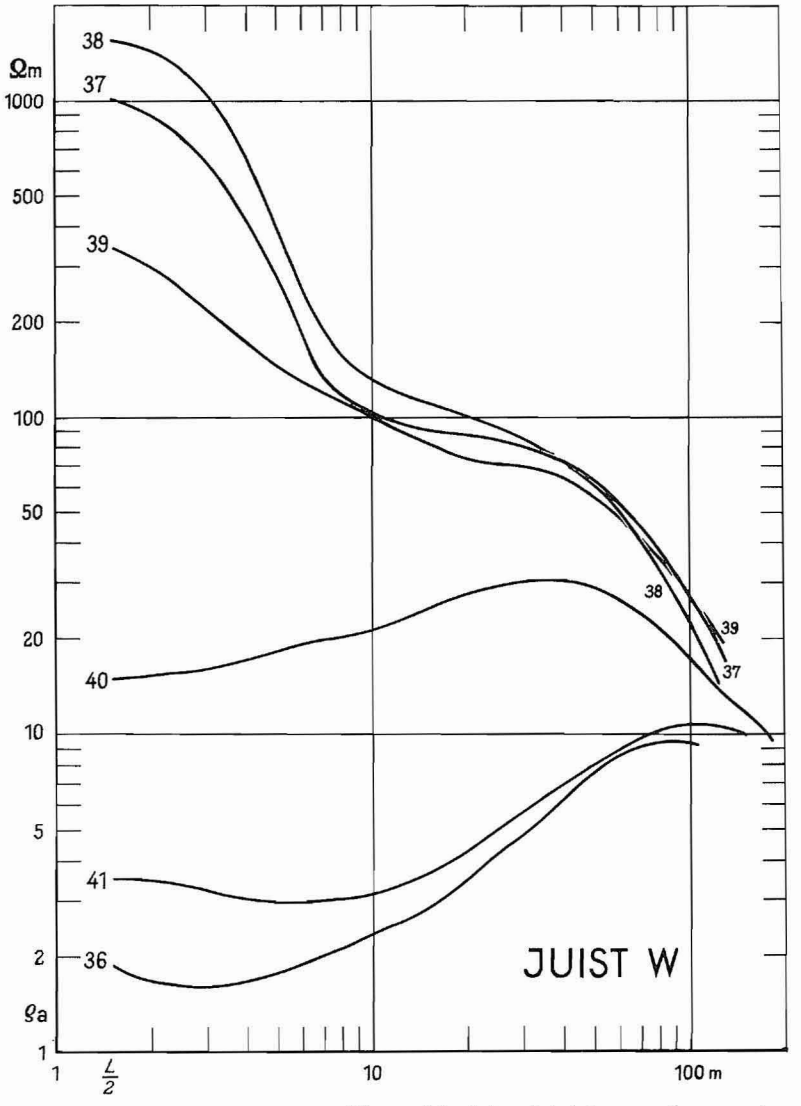


Figure 11. Isle of Juist; sounding graphs.

Hitherto, the seaside resort has been supplied from a freshwater lens immediately east of the village (Figure 10). In 1962, the extension of this lens was estimated by geoelectrical measurements. In addition, geophysics succeeded in determining and limiting a major lens at the western end of the isle (Bill). Parts of the results are given in the two vertical sections in Figure 10. The measured graphs of profile VI are plotted in Figure 11 which partly can be interpreted as simple two-layer cases. If the remaining part of the curve of the parallel profiles is also considered, the presence of an impermeable layer is indicated at approximately 20 m depth which is superposed by the freshwater lens. Otherwise the lense should be thicker at the measuring point 38 than at the neighbouring stations 37 and 39. On the other hand, at measuring point 35 (profile VIII) the boulder clay is presumably absent. According to the geology it can be presupposed that the layer of boulder clay, in view of the petrogenesis, cannot have a relief: at a certain depth it is either present or has been eroded. In the eastern part of profile VIII, a clay layer occurs at a few meters' depth (7 ohmm) which prevents an infiltration of the precipitation. Quite obviously, this layer developed from a bipartition of the isle which historically can be dated. On top of this clay layer, the freshwater lake as shown on the site plan, has formed: it is the rainwater that cannot infiltrate the clay layer. Furthermore, the peculiar delimitation of the freshwater lense in profile VI (stations 36, 40, 41) is a consequence of a historical event. Shortly prior to the geoelectrical measurements, parts of the isle were inundated owing to a hurricane tidal wave disaster. The penetrating sea water destroyed the marginal parts of the freshwater lense. An equilibrium based on the specific weight had not yet existed at the time of the measurements. Thus, in 1962, salt water was superposing brackish water at station 41.

The investigations on the Isle of Juist have showed that the geophysicist, in the interpretation of his measuring data and because of the hydrogeological situation, sometimes has also to consider the most recent history in addition to water chemism and petrography. It is a fact that the freshwater/saltwater interface is a fluctuating border which often reflects the historical development of the coastal area.

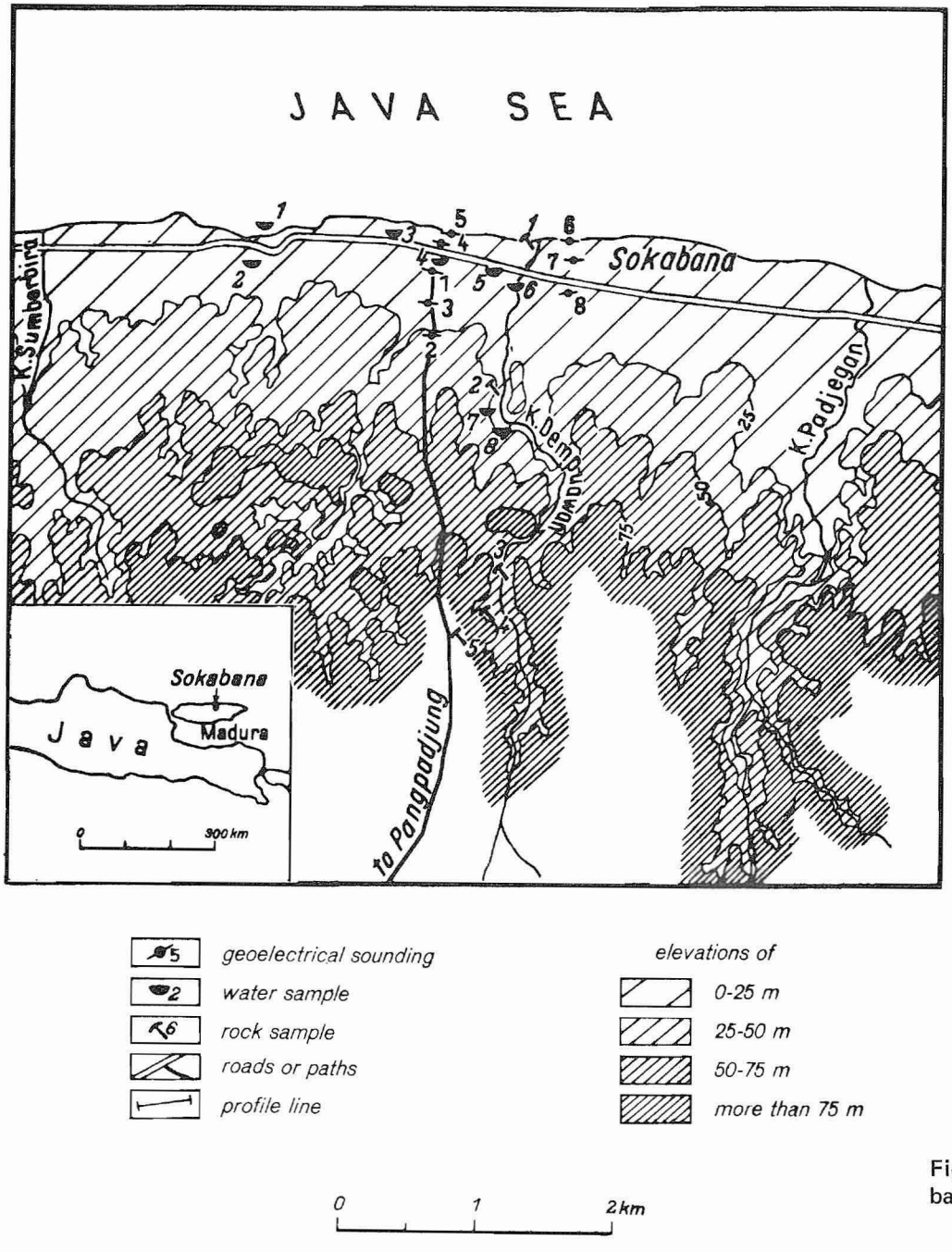


Figure 12. Investigation area of Sokabana, Madura.

Freshwater/saltwater interface in solid rocks, Sokabana (Madura), Indonesia. This example has as its objective the determination of the distribution of fresh and salty groundwater in solid rocks. It can be compared with the Isle of Juist where, apart from the chemism of the water (fresh, brackish, salty), the petrography of the loose sediments (sand, boulder, clay) determines the electrical resistivity distribution in the underground of the isle (Flathe and Pfeiffer, 1963).

Along the northern coast of the Isle of Madura, which is northeast of Surabaja in front of the eastern part of the Isle of Java, limestone formations plunge northeast into the Java sea (see Figure 12). Wells located there have proved that sufficient amounts of water can be developed to satisfy local demands. Naturally it is fresh water that is wanted. Since the water bearing parts of these limestones were possibly of only small thickness and it is known that from neighbouring valley sections the underlying bed consists of marly limestones unsuitable for water development, the geophysicist was confronted with the problem of clearing up the situation of the interface between fractured and marly limestones, and the situation of the interface between useful fresh water and useless salt water.

A relatively exact hydrogeological interpretation of measurements would not have been possible without a number of drillings present there, in spite of the resistivity variations covering almost 4 decimal powers in areas of less than 1 km extension (see sounding graphs Figure 14). Figure 12 shows a site

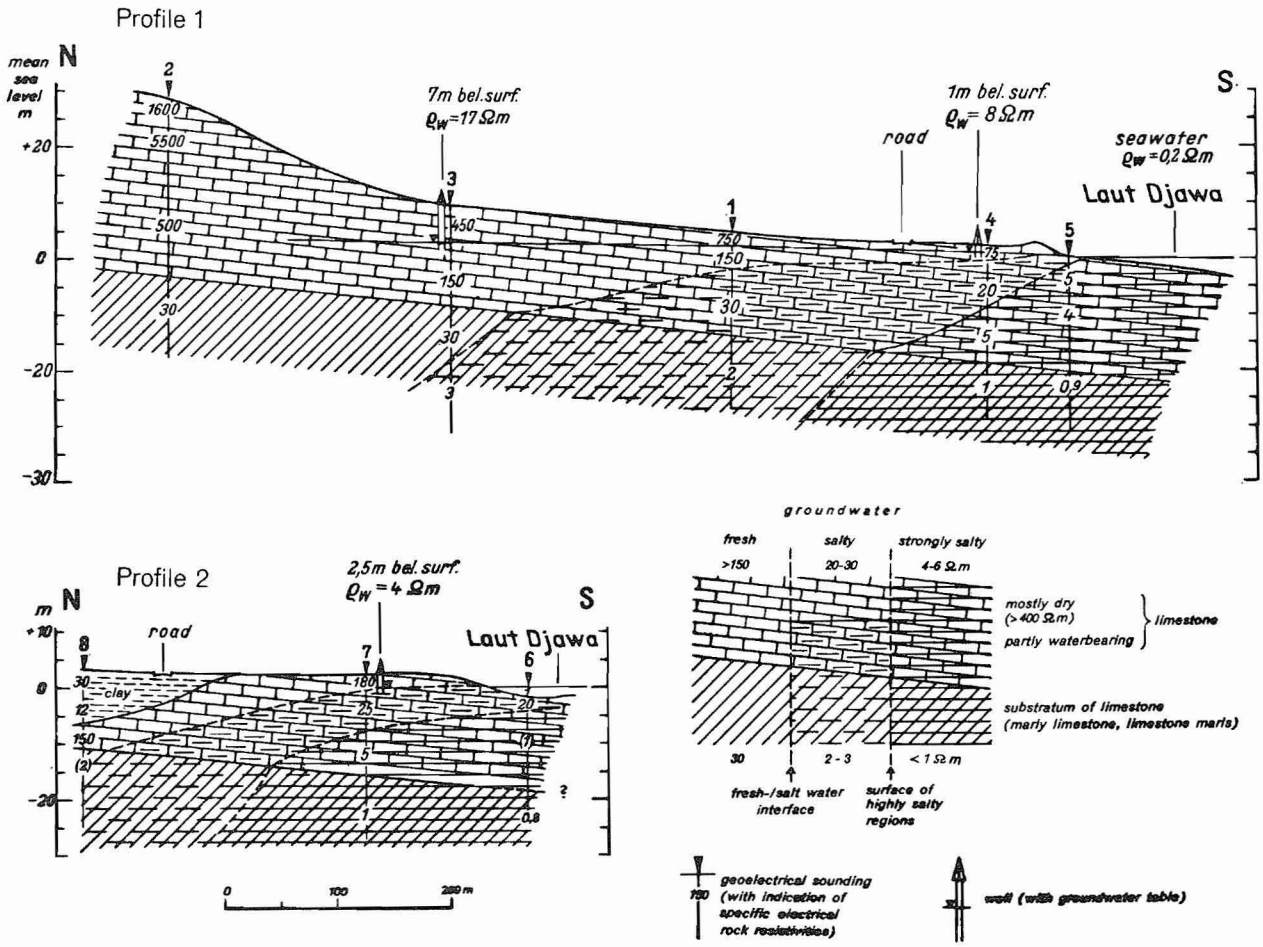


Figure 13. Vertical sections near Sokabana at the northern coast of the Isle of Madura based on geoelectrical investigations.

plan of the combined geoelectrical-hydrogeologic investigations, and Figure 13 the result in two vertical sections. The present wells gave evidence of the depth of the groundwater level, and the resistivities of groundwater in the wells.

These values plotted in the vertical sections facilitate a differentiation of the resistivity distribution that formed a concept for interpretation of the measuring curves represented in Figure 14. In other words, the hydrogeological knowledge obtained from present exposures narrowed down the principle of equivalence of the mathematical-physical interpretation by estimating certain ranges of resistivity for the intermediate layers. The picture shown in Figure 13 was drafted under consideration of the hydraulic behaviour in the superposition of fresh water above salt water because of the different specific weight of the waters for one, and the different permeability of compact rocks for the other. The legend below on the right gives the concept with which the hydrogeological interpretation was tackled. The profiles illustrate what may be one possible solution to the problem that would do justice to the given geophysical-hydrogeological data. An examination in detail may be left to the reader. Here equivalent solutions are possible. Looked at more closely, Sokabana represents a very attractive example for utilizing all possibilities for the combination of hydrogeological and geophysical investigations.

Groundwater salinity inland.

Salt water in humid areas with arid character, Chaco Boreal in Paraguay, South America. In 1959, the Bundesanstalt für Bodenforschung (Geological Survey) in Hannover sent off a team of specialists to carry out an extensive investigation of the Chaco Boreal of Paraguay. This investigation consisted of a combination of a hydrogeological inventory survey, a geophysical prospection and soil investigations for improving the agricultural utilization of the central Chaco. Hydrogeologically, this meant an analysis of the groundwater balance in an alluvial plain of more than 1000 km width that extends from the Bolivian Andes in the west to the Rio Paraguay in the east of the hydrostatic base level in a very gentle gradient of 30 m per 100 km. The substratum of the Chaco consists of fine-grained Quaternary and Tertiary loose sediments (grain size maximum 1 mm) of unknown thickness. Looking at the petrogenesis it was to be expected that the groundwater in the deeper underground would be salt-bearing, a fact that was fully confirmed by geoelectrical investigations. Of real interest, however, was a near-surface first groundwater layer with a thickness of some 10 m, of which the marly impermeable bed now and then showed a distinct relief. Locally infiltrated rainwater can easily collect here in pockets and channels, and with a carefully controlled pumping rate above underlying salt water (specifically heavier), it can be bailed out.

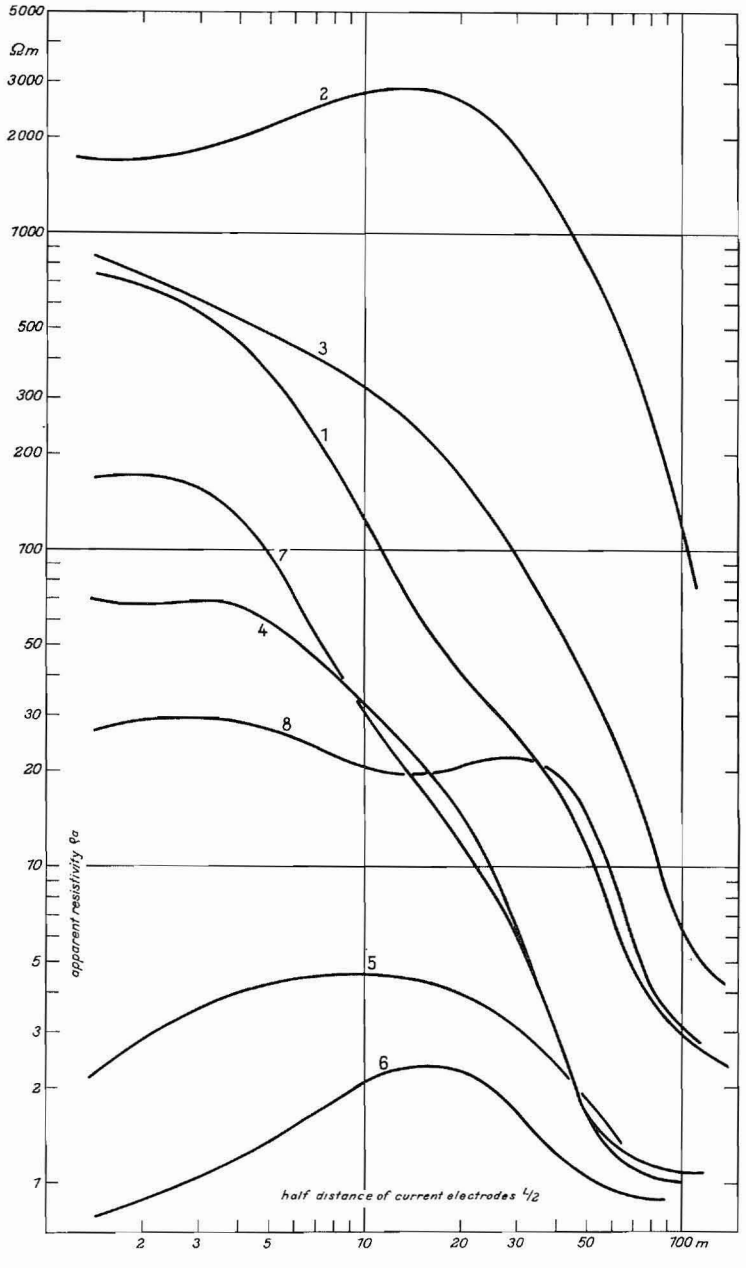


Figure 14. Geoelectrical sounding graphs recorded near Sokabana.

As an example, a geoelectrical detail investigation shall be described. It represents nothing more than a small part out of a very large and extensive total survey. The subject concerned is the water supply of Filadelfia (800 inhabitants) in the centre of a colony of Mennonites in the central part of the Chaco where the German language is spoken. This colony originated in the twenties when members of the sect of Mennonites coming from the USSR (Germans from the River Volga) and from Canada settled here. The problem for the geophysicist here is not only the determination of the interface fresh/salt water, but also – and that predominantly – the deepseated groundwater table. The whole ambiguity is illustrated in Figure 15, if the explanations made earlier are transferred to this objective. The caption under the picture may elucidate. Note the depths marked at each single sounding by thick bars. These depths alone can be determined by geoelectrical measurements. As far as fresh water is concerned, it is in no case the groundwater table, but either the interface sand/clay or the saltwater top.

Figure 16 shows yet another vertical section, with three wells which essentially contribute towards the interpretation of geoelectrical measurements (conjunction measurements). The four attached sounding graphs show, in one aspect, marked differences in resistivity, and for another, how complex in the present case the interpretation of physical results into hydrogeological facts may be.

Groundwater salinity in the southern Jordan Valley, Jordan. Consequent to intensive well-boring activity in the Jordan Valley, north of the Dead Sea, and to an increased drainage of groundwater for irrigation, the groundwater salinity in this region has in previous years become a serious problem. Further plans required an adequate knowledge of the position of fresh water bearing permeable loose sediments and their accurate delimitation against sediments in horizontal as well as in vertical direction (Figures 17 and 18).

The prospected graben section which is situated between the Palestinian and East Jordanian blocks, is filled up with post-Miocene sediments. In their deeper part, these sediments consist of an alternating bedding of claystones, sandstones and conglomerates. Superposing this series, which is estimated to have a thickness of roughly 300 m, are the Pleistocene 'Lisan Marls', fine-clastic sedimentations that towards the graben margins interfinger with coarser clastic erosion material coming from the graben margins. Alluvial fans at the larger wadi outlets partly advance into the graben centre and thin out in the Lisan Marls. Due to erosion the Lisan Marls are extensively dissected to form the impassable 'badlands' because of the structurally lowered position of the hydrostatic base level. These features determine, together with the erosionally less affected 'broken grounds', near the graben margin, the character of the 'Ghor', into which the actual Jordan flood plain is deeply cut known as 'Zor'. The maximum graben width, opposite Jericho, is about 25 km, the minimum width, near Karame, is 10 km.

This brief outline of the geology may suffice to understand the task with which geoelectrics was confronted.*

The technical logistics of taking measurements from November 1963 until February 1964 during an extremely rainy winter season, was rather difficult, especially in the badlands.

The very low specific resistivities of the Lisan Marls necessitated the measurement of very small voltages. A fully transistorized DC amplifier to be installed in a special field car, was constructed and developed by the Bundesanstalt für Bodenforschung (Geological Survey) in Hannover, specifically for application in arid and tropic regions.

The sounding graphs recorded in the Jordan Valley shown in Figure 19 are representative of the curves measured along the Jericho-Shuna section. They show a variation of rock resistivities

*Immediately following the geoelectrical investigations, F. Bender, *et al.* (1964) made a comprehensive hydrogeological study of the graben to serve as basis for the hydrogeological interpretation of results, with a view to future hydrological activities. They also prepared a geological map (scale 1: 50,000) of the southern Jordan Valley (Bender, *et al.*, 1964).

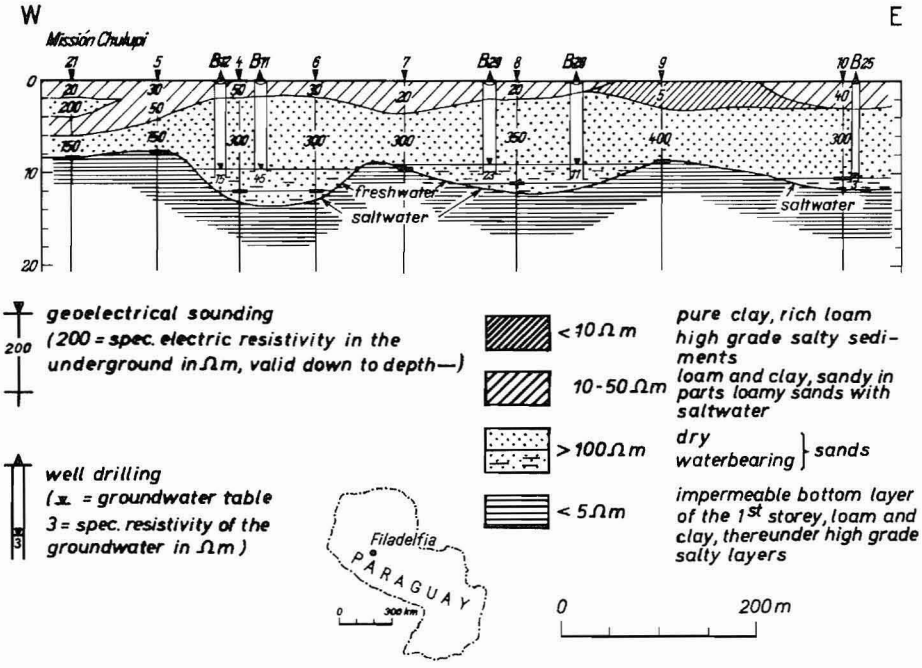


Figure 15. Vertical section from Filadelfia, Paraguay, central Chaco (Bender and Flathe, 1960). The specific electrical resistivities calculated from the sounding graphs are marked for the sands of the first storey as well as for the impermeable and semipermeable covering. A decrease of the electric resistivity indicates a decrease of permeability for water caused by loamy and clayey components which are good electric conductors. In the region of wells 11 and 12 the sands are very permeable (300Ωm) and the relatively thin cover is semipermeable (50Ωm). The basin at the bottom of the first storey contains fresh water underlain by salt water. Geoelectrically the fresh/saltwater interface was determined here. The difference in the C1 content of the waters from wells 11 and 12, reflected in the specific electrical resistivities of the waters (45Ωm and 15Ωm), results from the different depths of the wells. The same is valid for wells 28 and 29.

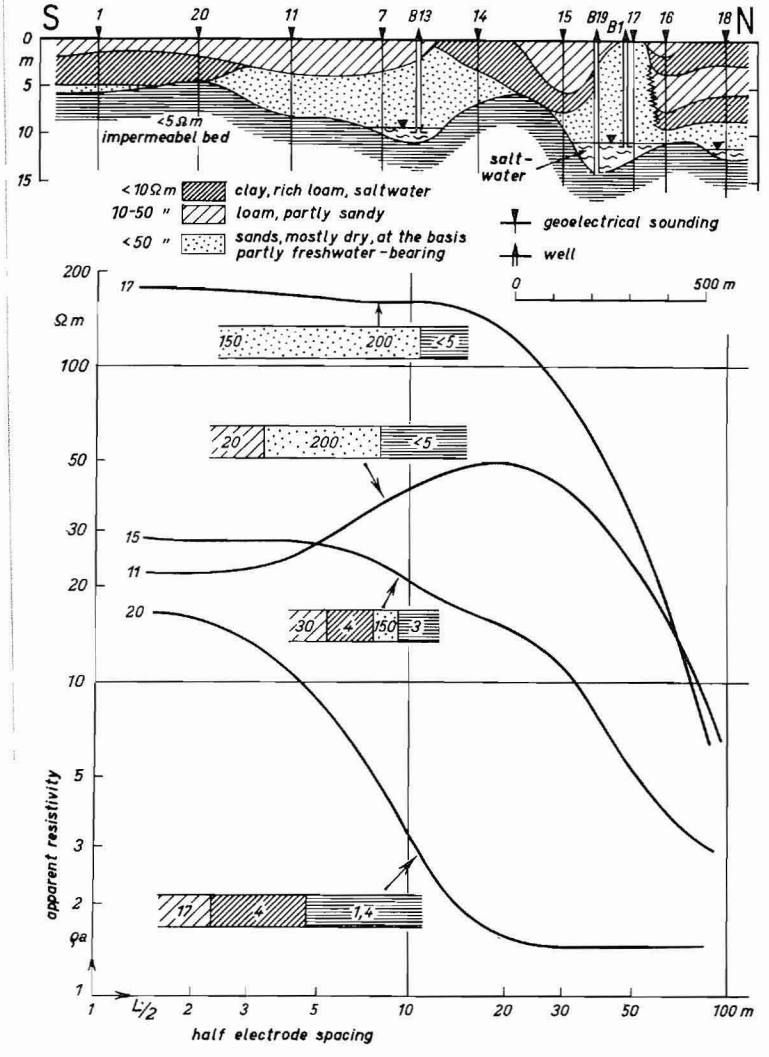


Figure 16. Filadelfia, Paraguay, vertical section and sounding graphs. Level of the saltwater table was detected in different depths, since the pockets are separated because of the relief of impermeable substratum.

covering nearly 4 decimal powers between 0,2 ohmm and 800 ohmm. The mathematical-physical interpretation by electronically computed standard graphs (see Figure 20) yielded a coherent picture of the true specific rock resistivities and thicknesses of layers or series of strata, respectively, in the underground, as far as these differ sufficiently in their specific resistivities. A purely physical interpretation based on the so-called principle of equivalence of geoelectrics cannot furnish an entirely unique picture. However, the equivalent possibilities for interpretation can substantially be narrowed down by means of conjunction measurements near drillings such as have been performed on the Jericho Observation Well (near S52), or by taking into consideration drilling results obtained in the more distant areas from the measuring points. This was, for example, done at the South Shuna Observation Well. In any case the interpretation has to be tackled by a hydrogeological concept determined by the respective objective. In this connection the basic reflections made earlier are referred to.

The horizontal section in 400 m below sea level, which is approximately the level of the Dead Sea, gives for the vertical Jericho-Shuna section an idea of the resistivity distribution in the underground (Figure 18).

At first glance one observes in the centre of the graben a zone with extremely low resistivities with 0,5 ohmm on the average, which is superposed on both sides by zones with resistivities of about 5 ohmm, that is a decimal power higher than in the centre. The interface of both of these zones generally dips into the direction of the graben border.* The smooth section plotted on

*They show a section typical for saltwater/freshwater interfaces.

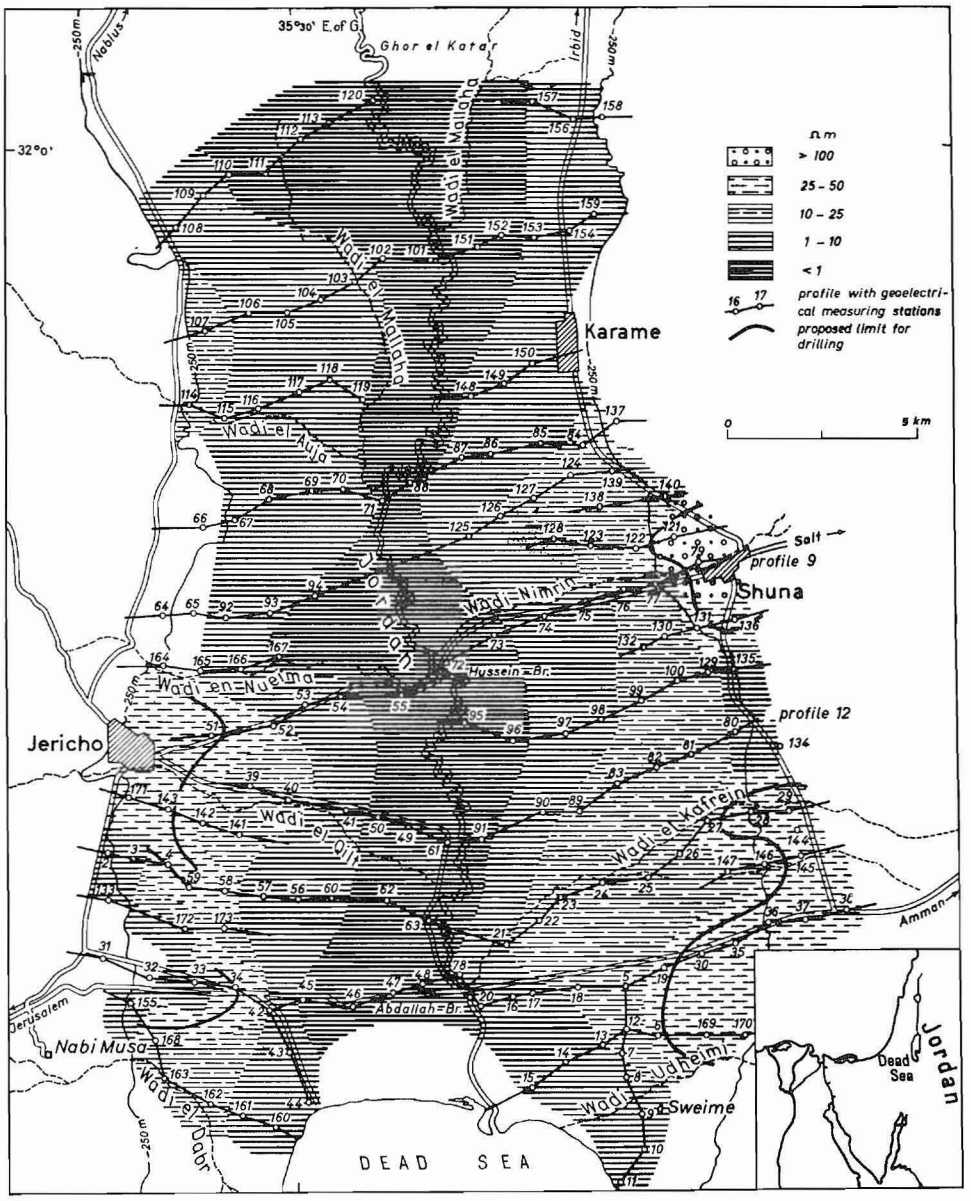


Figure 17. Jordan Valley, horizontal section with true specific resistivities in 400 m below sea level (Flathe, 1968).

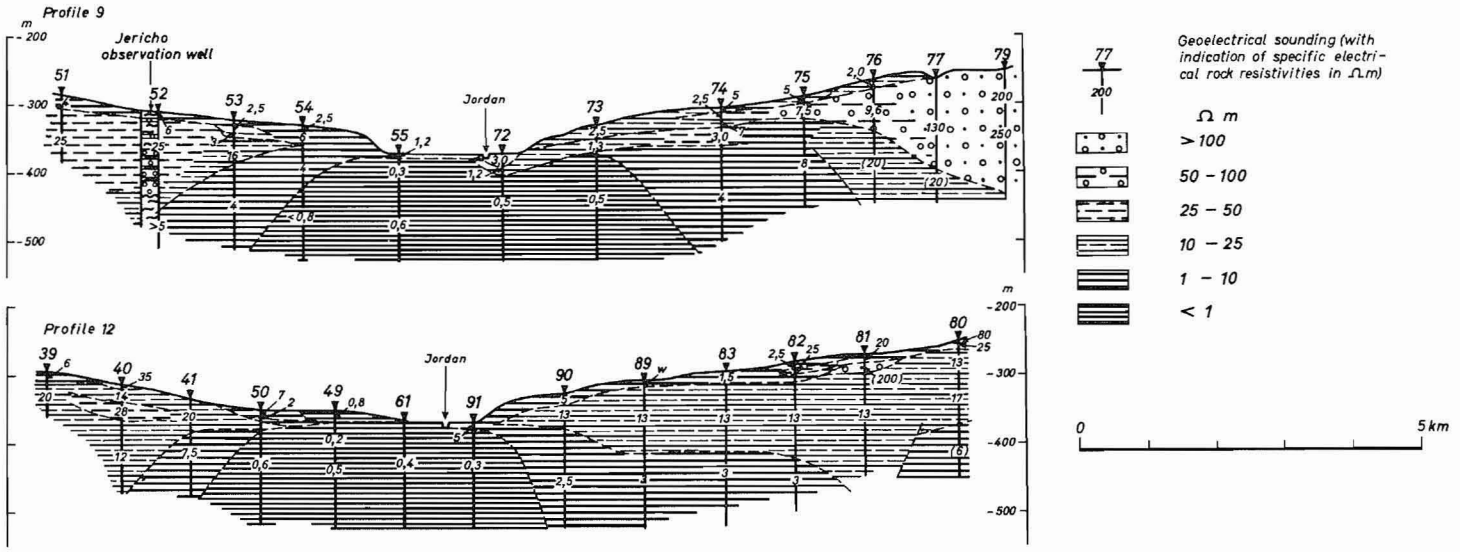


Figure 18. Vertical sections across the Jordon Valley.

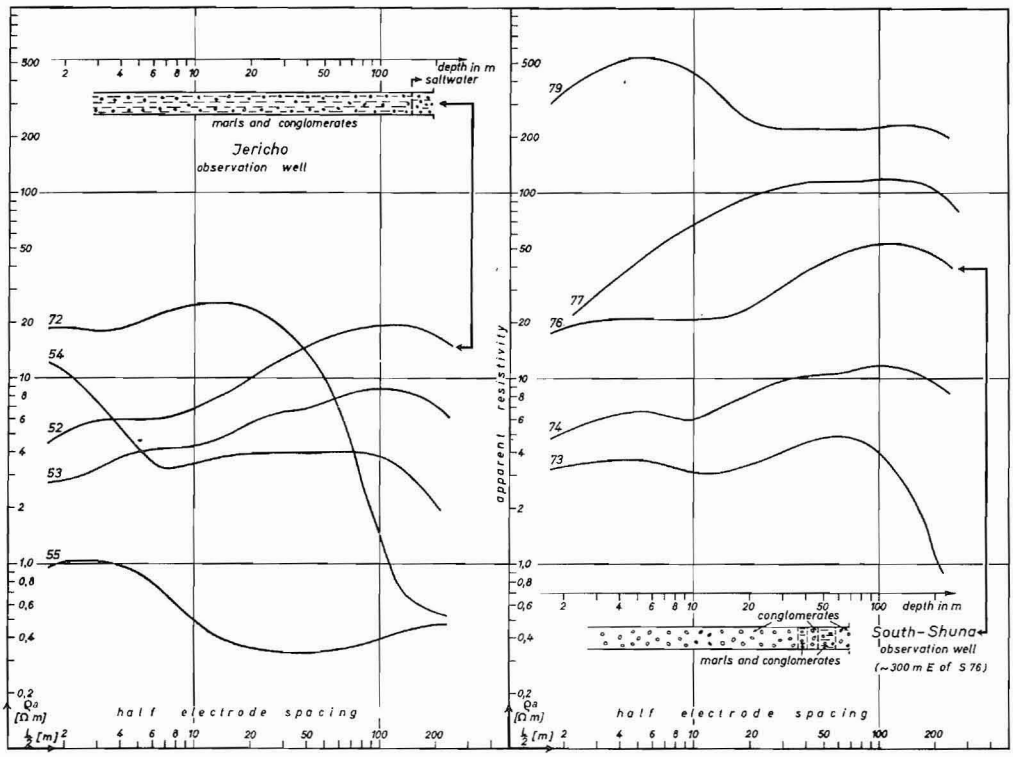


Figure 19. Sounding graphs recorded along the profile 9, Jericho – Shuna.

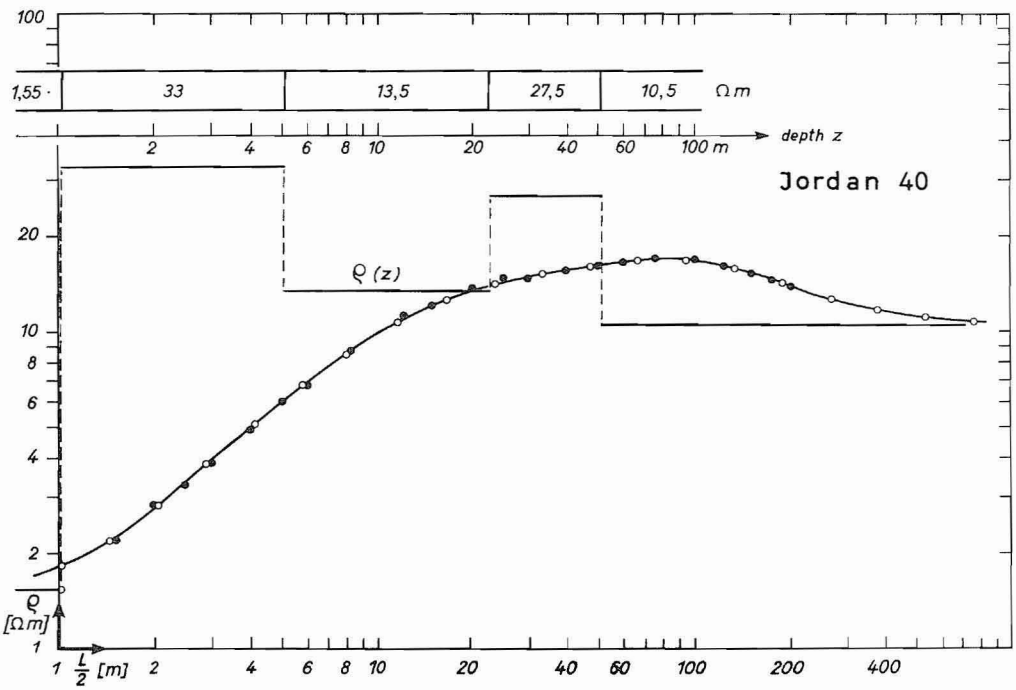


Figure 20. Comparison between measured graph and electronically calculated master curve (● recorded, ○ computed data).

the basis of an interpolation between measuring points with large intervals between, is a simplified representation of a complicated structure. Drillings have shown that the alternating bedding of permeable and impermeable beds is connected with a corresponding interfingering of pore waters with different electrolytic content, perhaps only because of the hydraulic conditions in the individual horizons that are differing from each other.

The heterogeneous structure in the centre of the graben – the conditions shown in the section reappear throughout the prospected graben area – suggests that sediments can be divided into two zones, based on the different salinity of the pore waters: one with extremely high salinity along the graben axis, and one strongly salted adjoining on either side. It is striking that a continuous transition is imperceptible.

On the other hand, a continuous increase of resistivities in the

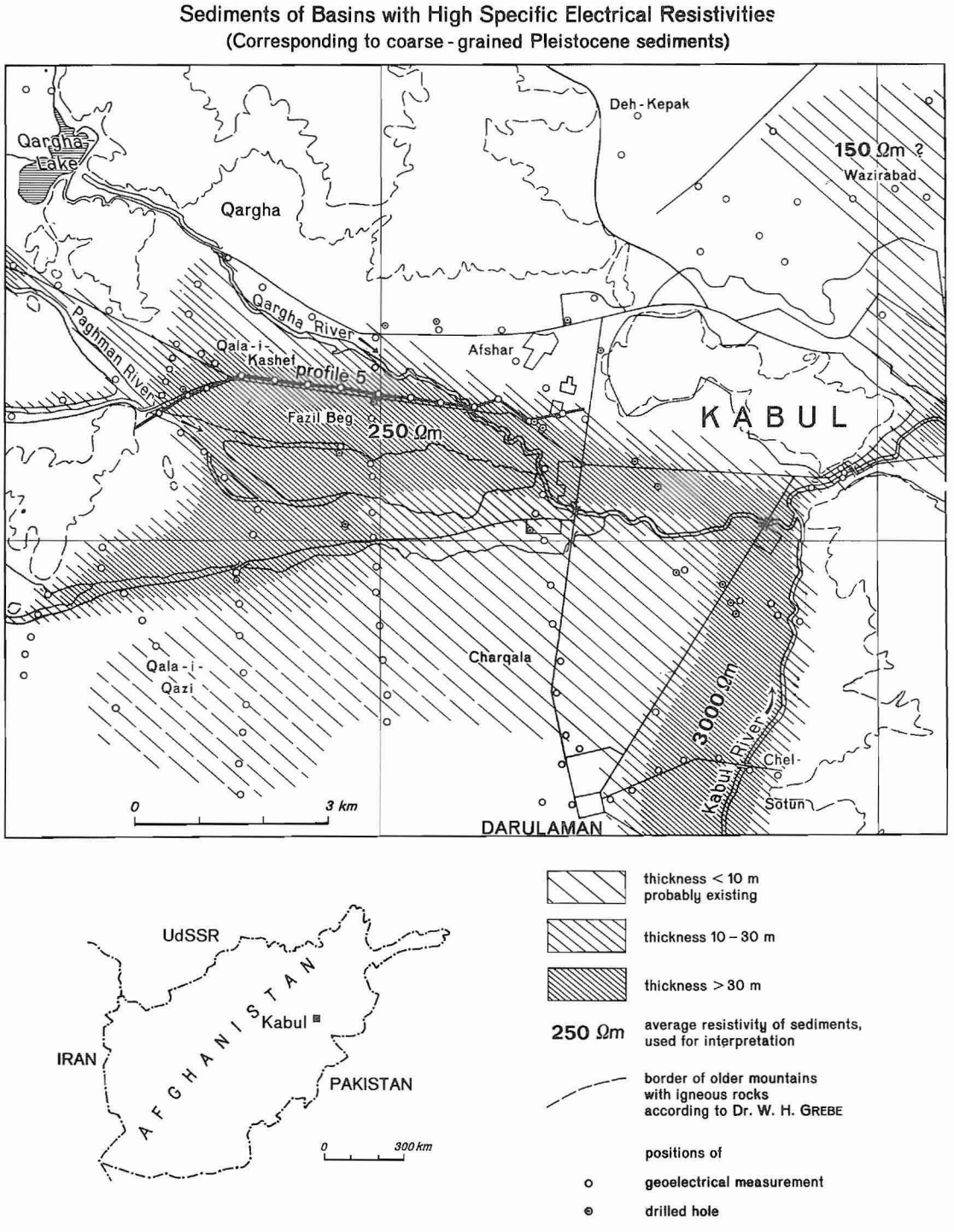


Figure 21. Aquifers in the Upper Kabul Basin (after Homilius, 1967).

deeper underground, in the direction of the graben border, appears outside the highly salted zones. The reason for this feature may be the decrease in the portion of electrically well-conducting clay horizons in the sediment series under simultaneous increase in the total thickness of permeable horizons with less salted waters. Thus, petrography and groundwater chemism determine the electrical resistivity, but neither of these influences can be divided by geoelectrical measurements.

Of special interest are the zones near the graben borders. In these areas, the vertical sections differ considerably. At the eastern graben edge, the profile No. 9 shows high resistivities down to depths far more than 100 m (S77, 79), whereas farther south in profile No. 12, apart from the overburden, the underground nowhere reaches resistivities exceeding 20 ohmm.

High resistivities in loose sediments indicate freshwater-saturated disintegrated rocks free of clay. Looking at the map in Figure 17, it becomes obvious that the older alluvial

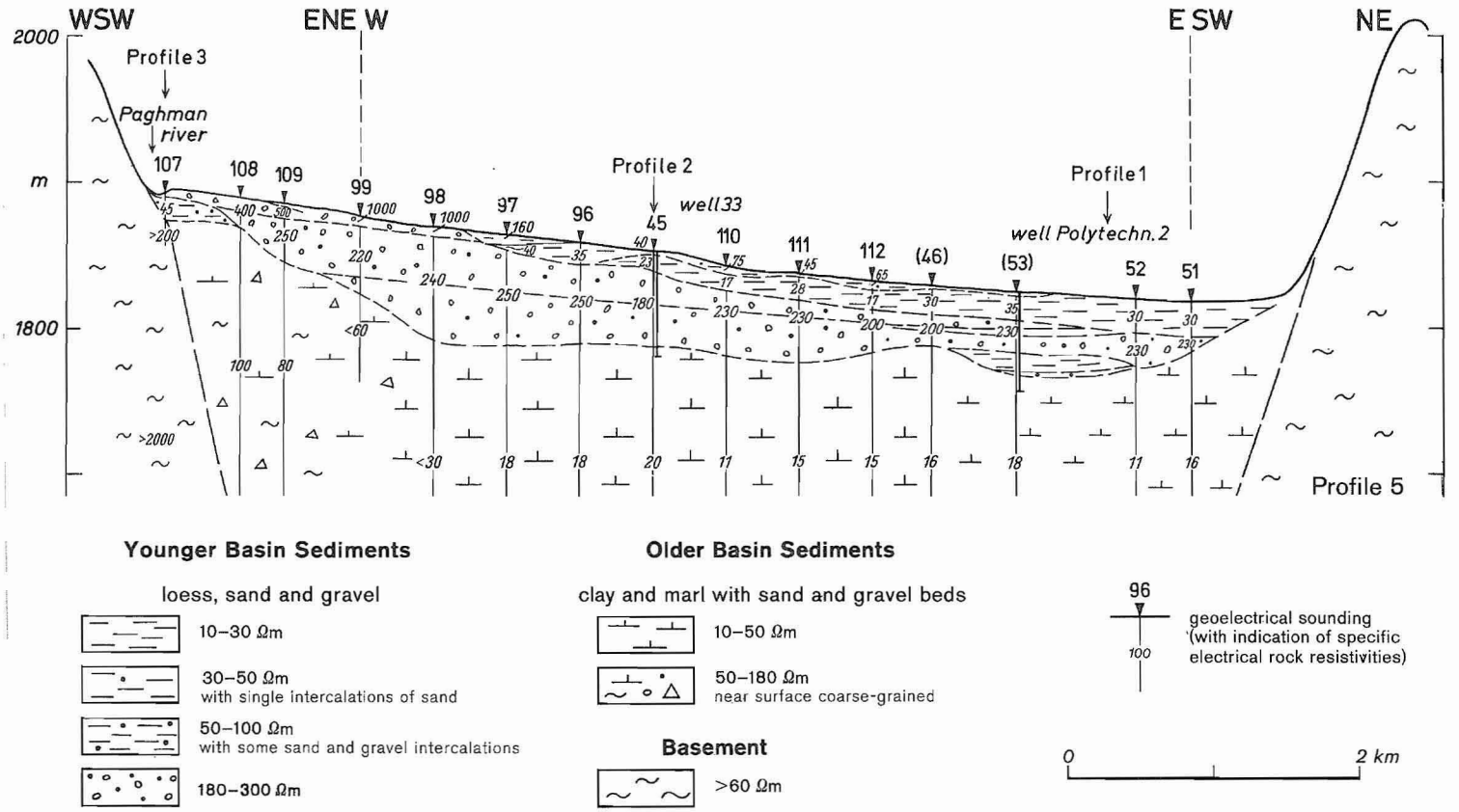


Figure 22. Vertical section in the Upper Kabul Basin (after Grebe and Homilius, 1968).

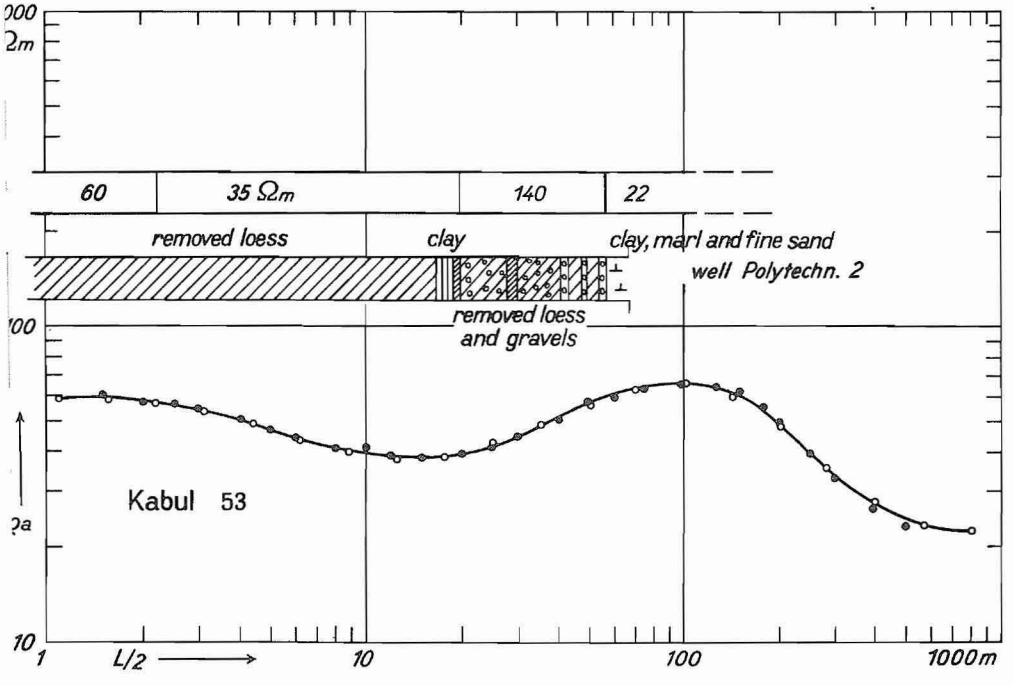


Figure 23. Comparison of geoelectrical sounding graph, drilling result and theoretically calculated master curve (● recorded, ○ computed, after Deppermann and Homilius, 1965).

fan in profile 9 which owes its origin to an aggradation coming from a lateral valley (Wadi Nimrin) is nowadays fed with fresh water coming from just this valley. This fresh water pushes the salt water towards the graben centre.

Conditions are similar at the western graben edge, but nowhere were resistivities found exceeding 100 ohmm. As evident from the Jericho Observation Well, impermeable clayey-marly sediments prevail.

Assuming a resistivity of 25 ohmm as the lower limit for a series of strata which from the water-economic point of view is

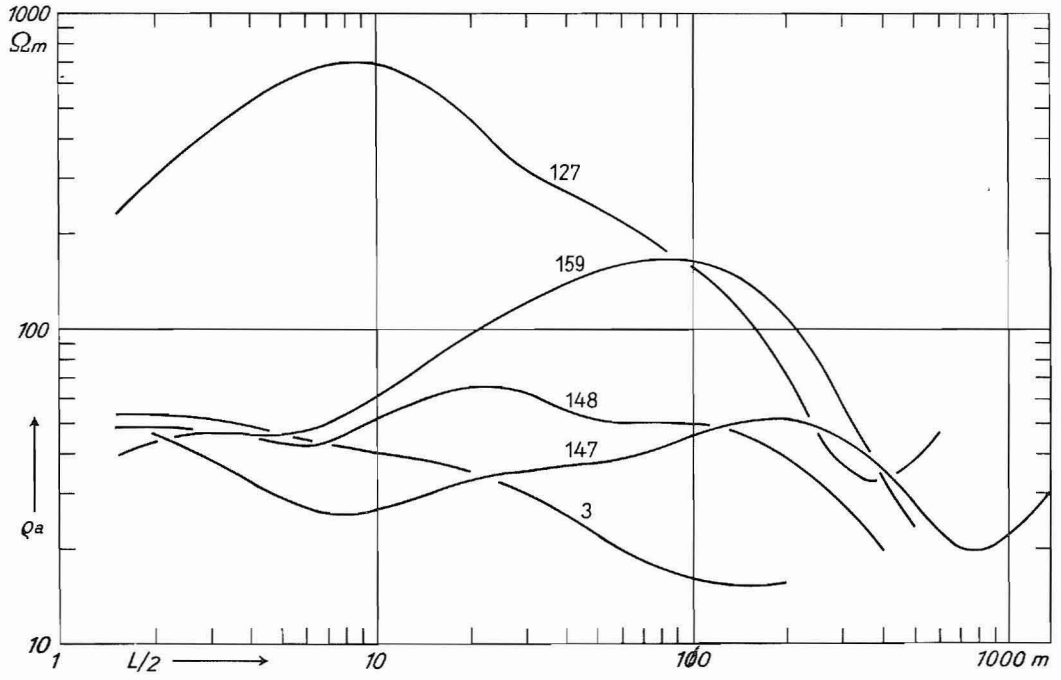


Figure 24. Typical sounding graphs from the Upper Kabul Basin (after Homilius, 1967).

water-promising, we have, under consideration of a safety margin, areas of interest for well borings. In Figure 17, such areas are marked. Drillings can be carried out in these areas down to the respective levels, with chances for obtaining useful groundwater.

A detailed study of the results constitutes the basis for further water development projects. One must keep in mind that the investigations performed in 1963-64 more or less represent a 'snapshot'. Any further water development would mean an interference in the total water balance which is subject to annual variations, anyway. Thus it would change the picture provided by geophysical investigations, of the groundwater salinity in the southern Jordan Valley.

Search for and delimitation of an aquifer.

Upper Kabul Basin, Afghanistan. The following case was chosen from a multitude of projects, as a practical example of the search for and delimitation of sandy and gravelly sediments for water development. In 1965-67, within Technical Assistance for Developing Countries, Homilius (1967) carried out extensive geoelectrical measurements in Afghanistan. The surroundings near the capital Kabul were investigated to ascertain the town water supply. Attempts to develop hitherto unknown aquifers by geophysical measurements proved successful.

In the Upper Kabul Basin, a total of 101 geoelectrical depth soundings were carried out. The hydrogeological situation is determined by the three rivers, Kabul, Paghman and Qargha flowing into the basin. Figure 21 shows the distribution of the sandy-gravelly sediments determined by geoelectrical measurements and considered suitable for water development.

The vertical section given in Figure 22 gives information on thickness and depth range. The interpretation of the sounding graphs is in this case based on a conjunction measurement (S53) at the well near the Polytechnic School. For this sounding graph, a standard graph was computed (Figure 23). The vertical section given in Figure 22 can be compared with the example of the valley cross section described in the first part of this paper.

In the Upper Kabul Basin there is an aquifer situated below a partly sandy, partly clayey overburden. This is expressed in the curve diagram. The measuring graphs plotted in Figure 24 from the Upper Kabul Basin can be compared with the curves in Figure 5.

From an ascent of the rear curve branches in S147 (L = 3 km) and S148, the basement can clearly be recognized. For water development, this result is certainly of no interest. Nevertheless it contributes to the clarification of the geological situation as a whole.

This example of the most recent measurements shall conclude this paper. It is left to the reader to draw comparisons between the principles outlined earlier and the case histories.

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Borehole geophysics as applied to groundwater

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Abstract. The Water Resources Division of the U.S. Geological Survey is conducting research on the application of borehole geophysics to groundwater investigations. The various logging techniques include: natural gamma, gamma-gamma, neutron-gamma, neutron-neutron, spontaneous potential, single point and short and long normal resistivity, caliper, sonic velocity, flowmeter, radioactive tracer, fluid resistivity, and gradient and differential temperature. An inexpensive system for recording logs on magnetic tape also has been developed. Quantitative interpretation of these logs will provide numerical values for the elements necessary to construct analog models of groundwater systems and data from logs is of considerable value in the design and interpretation of surface geophysical surveys. Geophysical logs can be interpreted in terms of the geometry, resistivity, bulk density, porosity, permeability, moisture content, and specific yield of water-bearing rocks and the source, movement, and chemical and physical characteristics of water.

Information obtained from interpretation of geophysical logs is usually more objective and consistent than that from an examination of cuttings or cores. In addition, geophysical logs are often the only means of obtaining information from old wells and lateral extrapolation of geologic and hydrologic conditions. When log interpretation factors have been established for a given environment, logging can partially supplant coring. Data obtained from samples become more meaningful when used with geophysical logs because logs provide continuous average values for a volume of material much larger than that of extracted samples. To place values on log response the geophysical equipment must be properly calibrated and standardized, and correction factors applied to nullify extraneous effects. New, lightweight logging equipment is economically compatible with present well costs and with the value of water. Using this type of equipment and the proper interpretative techniques, borehole geophysics will become a significant tool in the development of the world's groundwater resources.

Résumé. La Division des ressources hydrauliques de la Commission géologique des E.-U. mène une recherche sur l'application de la géophysique des sondages à la recherche des eaux souterraines. Les diverses techniques diagraphiques comprennent les méthodes suivantes; le gamma naturel, le gamma-gamma, le neutron-gamma, le neutron-neutron, le potentiel spontané, le carottage, électrique à point unique, à longue et courte séparation, le compas de précision, la vélocité du son, le fluxmètre, le traceur radioactif, la résistivité des fluides ainsi que la température de gradient et la température différentielle. Un système peu coûteux d'inscrire les mesures sur ruban magnétiques a également été mis au point. L'interprétation quantitative de ces diagrammes fournira des valeurs numériques pour les éléments nécessaires à la construction de maquettes analogiques de réseaux des eaux souterraines; et les données des diagrammes sont d'une valeur inestimable pour la conception et l'interprétation des relevés géophysiques de surface. Les diagrammes géophysiques peuvent être interprétés en fonction de la géométrie, de la résistivité, de la densité globale, de la porosité, de la perméabilité, de la teneur en humidité et du débit spécifique des roches aquifères de même que de la source, du mouvement et des caractéristiques chimiques et physiques de l'eau.

Les renseignements obtenus de l'interprétation des diagrammes géophysiques sont d'ordinaire plus objectifs et plus uniformes que ceux qui proviennent de coupes ou de carottes. En outre, les diagrammes géophysiques constituent souvent le seul moyen d'obtenir des informations des vieux puits et ils facilitent l'extrapolation latérale des conditions géologiques et hydrologiques. Lorsque les facteurs d'interprétation du diagramme ont été établis pour un milieu donné, la diagraphie peut remplacer partiellement la carottage. Les données obtenues d'échantillons sont plus révélatrices lorsqu'elles sont utilisées concurremment avec les diagrammes géophysiques parce que les diagrammes fournissent des valeurs moyennes continues pour un volume de matériau bien plus considérable que celui des échantillons extraits. Pour permettre d'évaluer la réponse du diagramme, l'équipement géophysique doit être étalonné et normalisé de façon appropriée et les facteurs de correction doivent être appliqués pour annuler les effets parasites.

Un équipement de diagraphie nouveau et léger est économiquement compatible avec les coûts actuels des puits et la valeur de l'eau. Grâce à ce genre d'équipement et aux techniques appropriées d'interprétation, la géophysique des sondages deviendra un outil très utile dans l'exploitation des ressources mondiales d'eau souterraine.

Exploratory drill holes or wells are our only means of direct access to the groundwater environment for information and fluid production or disposal. Drill holes are an expensive means of access to the lithosphere and techniques such as well logging are economically justified because they enable us to derive more data from each hole and extrapolate this information laterally. Borehole geophysics or well logging includes all techniques of lowering a sensing device in a well to make a record which may be interpreted in terms of the rock characteristics, contained fluids, and the construction of the well. Although the main subject of this paper is the utilization of borehole geophysics in groundwater investigations, similar applications may be made in other fields of mineral development. More holes likely are drilled for water than for all other purposes combined. Each of these holes is a valuable sample of the geologic environment and information derived from them can aid in the development of water supplies, sand and gravel deposits, other nonmetallic and metallic mineral deposits, petroleum, underground waste disposal, and storage sites for liquids and gases, and provide engineering data necessary for construction.

The petroleum industry is responsible for the development of almost all of the logging equipment and interpretive techniques used today. It is unlikely that borehole geophysics would have reached its present state of development without the economic impetus provided by the search for oil. The world-wide search for groundwater may never reach this economic level but we can take advantage of the existing techniques paid for by oil. A thorough knowledge of developments in petroleum well logging is essential to avoid duplication of effort in basic logging research in the field of water resources. The greatest need now is for research on the

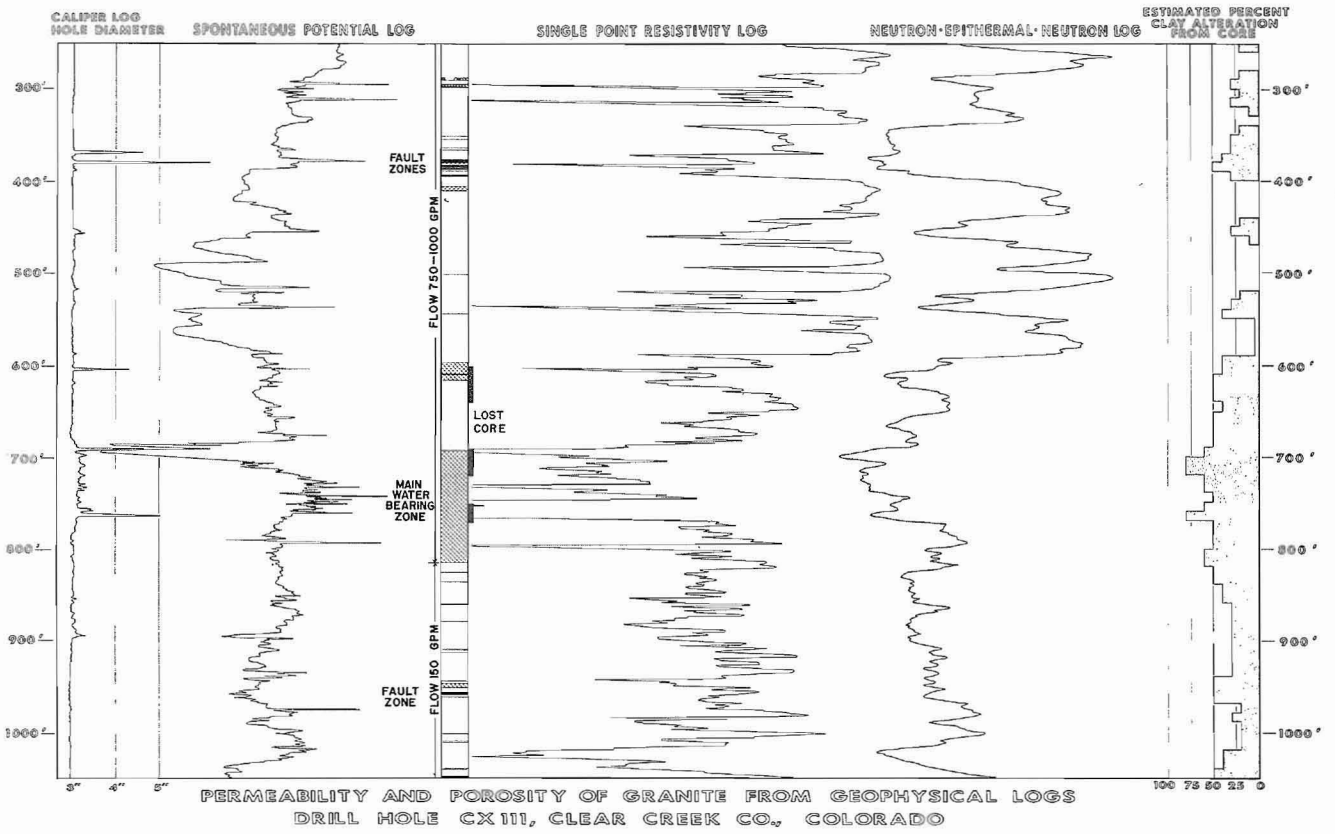


Figure 1. A comparison of geophysical logs with hydrological and geological data obtained during drilling and packer testing (left center) and clay alteration based on core (far right).

application of petroleum logging techniques to other fields of mineral exploration and modification to bring equipment and service costs in line with the value of the product.

The Water Resources Division of the U.S. Geological Survey is conducting research in borehole geophysics as applied to geohydrology. A one- to two-man research project has been under way for 4 years with headquarters at the Denver Federal Center, Denver, Colorado. The project objectives are to investigate, appraise and report on the current and potential applications of qualitative and quantitative borehole geophysics in groundwater hydrology and to investigate the value of these techniques in the various groundwater environments throughout the United States. Other objectives are to develop and modify geophysical logging equipment as an aid in the solution of specific water problems and to improve and adapt methods of quantitative and qualitative log interpretation to groundwater studies. A comprehensive discussion of all logging methods presently being used in groundwater hydrology or with potential application is beyond the scope of this paper. For this reason descriptions of equipment and specific applications of logging techniques in this report refer to the Water Resources Division research project.

Current application of logging in groundwater

The application of geophysical well logging to groundwater hydrology is comparable to its use in petroleum exploration in the 1930s. In 1964 approximately 436,000 new water wells were drilled in the United States; it is doubtful if more than 1 percent of these wells were logged with geophysical equipment (Meyer and Wyrick, 1966). The conventional driller's or geologist's log of cuttings is made on most water wells but these are not of consistent value. Properly made geophysical logs are unbiased and thus may be utilized over wide geographic areas and for long periods of time without loss of meaning.

The single most important factor in the present minor use of geophysical logging in water wells is cost. It is estimated that the annual investment in water wells in the United States is one half to three quarter billion dollars, not including pumps and plumbing. Although the volume of drilling and total money invested in groundwater development is very high, the unit value of water and water wells is almost never equal to the unit value of oil and oil wells. The cost of logging equipment and commercial logging services must be compatible with well costs and the value of the product. An estimated average cost for each United States water well is $1500; however, this reflects the preponderance of shallow, small-diameter domestic wells in the East. In the western United States expensive municipal, irrigation and disposal wells are comparable in cost to some oil wells. The general trend is toward deeper, more expensive wells, which will require, and will economically justify, more advanced knowledge of the groundwater environment and better well construction and completion practices. As a result of recent developments in electronics, small and relatively inexpensive logging units that can be operated by one man are now available. Equipment similar to that described in this report should improve the cost-benefit ratio for water well logging, and result in wider use of borehole geophysics in groundwater hydrology.

A second basic deterrent to the widespread use of geophysical well logs in groundwater is that few professional geologists,

hydrologists, or engineers are working on water well drilling projects. In addition, most drillers, municipal water engineers, or hydrologists do not have the experience necessary to interpret geophysical logs. Therefore interpretation must be provided along with logs, if logging is to have much practical application to the solution of hydrologic problems.

The many water wells being drilled throughout the world provide the sample points for an invaluable fund of geological knowledge if the proper data are obtained. Most geophysical logs now are interpreted with a single objective, such as water or oil, yet valuable mineral deposits have been discovered by logging water wells. For example, in Florida, gamma-logging of small diameter domestic water wells has yielded information valuable in exploration for phosphate deposits. The logs shown in Figure 1 were made for a hydrologic study. However a comparison of clay content of the core and the neutron log suggests that neutron logs may provide a means for measuring the degree of alteration of granite—a guide to ore deposits in some areas. In a few areas, water wells have been logged for stratigraphic or structural information to guide oil exploration. An important benefit to be derived from the increased logging of water wells will be additions to our geological knowledge of the subsurface which can guide the development of all types of mineral resources.

Log interpretation

In this age of specialization a professional log analyst might be regarded as a specialist by a hydrologist or geologist and as a generalist by a nuclear physicist or electronics engineer. To have a complete understanding of the theoretical basis for all of the logs described herein a log analyst trained in either geology, geophysics, or hydrology would have to be knowledgeable in at least the following additional fields: electronics, acoustics, nuclear physics, chemistry and higher mathematics. Fortunately, it is still possible to use logs without having all of this background. The theoretical basis for most logs is complex, particularly so, in equations dealing with the distribution and density of gamma photons or neutrons in the well bore and adjacent strata. Barsukov (1965) derives extremely complex equations for neutron density in the well and adjacent strata. The statement is then made that "the complexity of the expressions . . . makes an analysis of the physical meaning of the terms of these equations impossible."

The number of variables in most theoretical equations describing log response further limits their use. In many cases not enough is known about a hole or the rocks penetrated to make all of the corrections necessary for quantitative interpretation; however, it is still possible to make qualitative log interpretations by empirical methods.

Empirical methods, based on the relationship of log response to samples, core, models, or experimentally derived equations, are the most widely used methods of interpretation. Most geophysical logs do not have a unique response: a certain recorded deflection may be due to one lithology in one area and another elsewhere. However, within a given geohydrologic environment log response is usually consistent for the same rock types. In such areas log character can be related to physical parameters measured on core samples or to properly constructed borehole models. However, in all of these methods either borehole effects must be corrected, or deflections on the log due to nonstandard hole conditions must be eliminated from quantitative consideration.

Quantitative interpretation of geophysical logs greatly depends on accurate calibration and frequent standardization. Calibration includes techniques of relating the magnitude of log response to the values of such parameters as porosity, bulk density, resistivity, temperature, and vertical flow in gallons per minute. Calibration may be done in a hole where core samples have provided the required values, or in a model which approaches hole and environmental conditions. Values from such models should be recorded with values measured at the same time on a field standard. Almost all measuring systems exhibit drift and the frequent use of field standards provides a means of compensating for this problem.

Most logs except those measuring in-hole fluid characteristics are susceptible to shifts caused by changes in the character of the drill hole. These unwanted shifts are due to differences in the distance from the sensor to the rock to be measured and changes in character of intervening material such as borehole fluids and casing. Wherever possible, correction for borehole effects should be made before any quantitative log interpretation is attempted.

The radius or depth of investigation, or volume of influence, is here considered to include that part of the formation contributing 90 percent of the recorded signal. The depth of investigation varies from a few millimeters for microresistivity logging devices to many meters for focused resistivity sondes. The action of drilling and circulating fluids causes changes in hole diameter, the formation of mud cake, and fluid invasion of the rocks we want to measure. For this reason information on the character of the hole, drilling techniques used, and the radius of investigation of the various logs are most important to proper log interpretation. Several types of resistivity logging devices are used in petroleum exploration to measure the thickness of mud cake and invasion which are related to the formation permeability.

The detector for a natural gamma log is at the center of the volume of influence and the center for gamma-gamma and neutron logs lies somewhere between the detector and the radioactive source. Because the influence extends for some distance above and below the detector a rock layer evidently begins to effect the recorded radiation before the detector enters the bed and continues to effect the response after it leaves the bed. This concept explains why a log will not attain a value representative of the lithology of a bed unless that bed is thick enough to fill the volume of influence. The 'thin bed effect' is an important consideration in log interpretation in many environments. Some multielectrode resistivity logs exhibit reversals in thin beds. The fact that gamma-log response is related to both the concentration of natural radioisotopes present and thickness of the bed led to a new technique for log interpretation.

Previously, quantitative log analysis was largely based on the relationship between maximum amplitude of the curve and the desired lithologic parameter. Scott, *et al.*, (1961) show that the area under the gamma curve is proportional to the product of the bed thickness, and the amount of gamma-emitting isotopes present. For practical applications their equation is reduced to the following: $\overline{G}_\gamma T = kA$ where $\overline{G}_\gamma$ is the average radiometric concentration in the bed, T is the thickness of the layer, k is the constant of proportionality determined by instrument calibration and A is the corrected area under the gamma curve. This theory assumes that sedimentary rock consists of a number of layers

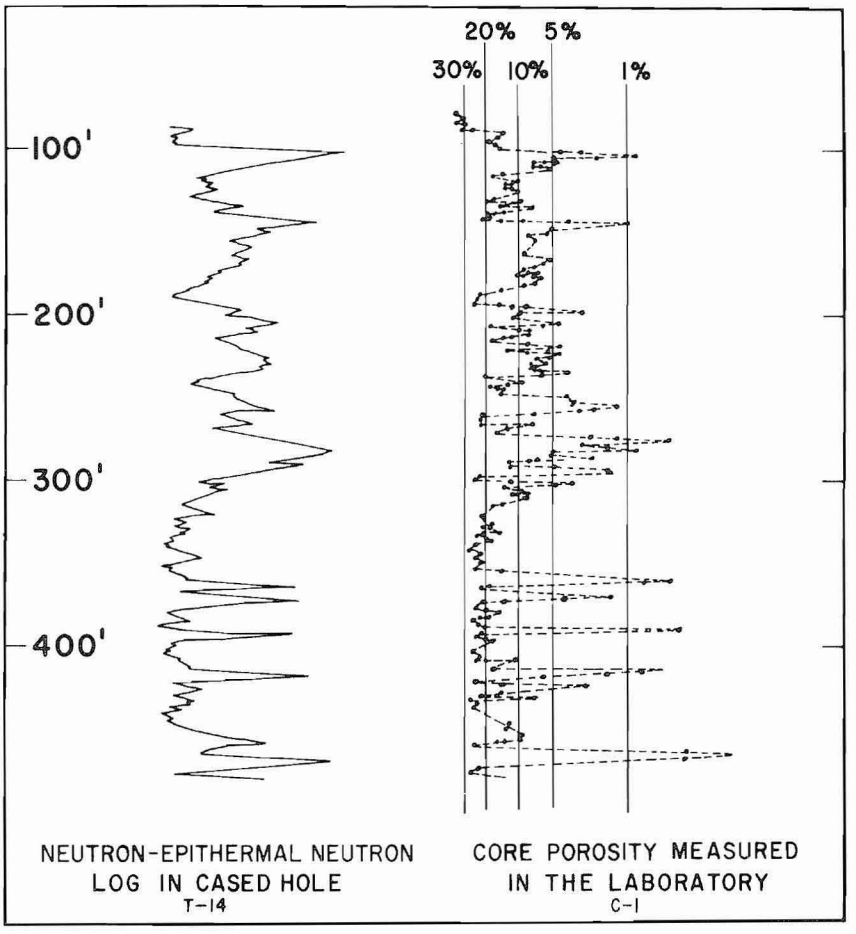


Figure 2. A comparison of a neutron log made in a cased hole with porosity measured on approximately 200 core samples obtained from an adjacent hole.

with differing concentrations of radioisotopes and that the recorded gamma log is a graphic summation of the curve representing each layer. This technique should be applied experimentally to other logs as it is likely to provide more accurate quantitative interpretation than the use of maximum amplitude, when bed thickness data are available.

Czubeck (1966) derives the theoretical relationship for using 'similitude' for quantitative interpretation of gamma-gamma logs. He suggests using artificial bulk density models to measure the changes in log response that occur when the spacing between the source and detector is changed. These spacing - count rate curves would then be related to variations of intensity in a drill hole measured with a constant spacing, but in an environment of variable bulk density.

Although core and cuttings from a logged hole provide one of the best and simplest ways of calibrating and correcting log response there are some problems in relating core values measured in the laboratory to log response. In the first place, measurements made on core are point samples; logs usually represent an average value of a sample much larger than the core extracted from the hole. New gamma attenuation equipment, which provides continuous measurement of core porosity, demonstrates the nonhomogeneous nature of some rocks (Evans, 1965). Under these conditions many samples should be analyzed to establish representative values. If the samples are representative the next problem is to relate the point value obtained from the sample to the log, which is a continuous record of averaged values. Unless core recovery is 100 percent the exact location of the sample is subject to error which may be very large in a single run with high core loss. The geophysical log also is subject to footage discrepancies due to cable stretch, equipment malfunction and human error. The usual technique, which can lead to significant errors, is to plot the samples on the same depth scale as the log and adjust them vertically until the best fit is obtained.

A recent paper describes the use of correlation functions, calculated manually or by computer, which can be used to derive the best relationship between samples and logs (Jeffries, 1966). The right side of Figure 1 shows how closely a record of average core values can compare to the response of a geophysical log. In this case the visual estimate of percent of clay alteration of granite was made before the neutron log. More detailed changes in alteration in the core undoubtedly could be detected using the neutron log as a guide to core examination. Figure 2 compares a neutron log with approximately 200 core samples analyzed in the laboratory. Note how much easier it is to interpret the log between 200 and 300 feet than the numerous core samples in this interval. The difficulty of relating the depth of one of the point samples to a given log value is also apparent. The common procedure of selecting a few pieces of core for laboratory analysis would have produced meaningless data in a sedimentary section exhibiting such inhomogeneities.

As a rule, the more different types of geophysical logs that are available for a single well and the more wells within a given geohydrologic environment that can be logged, the greater the

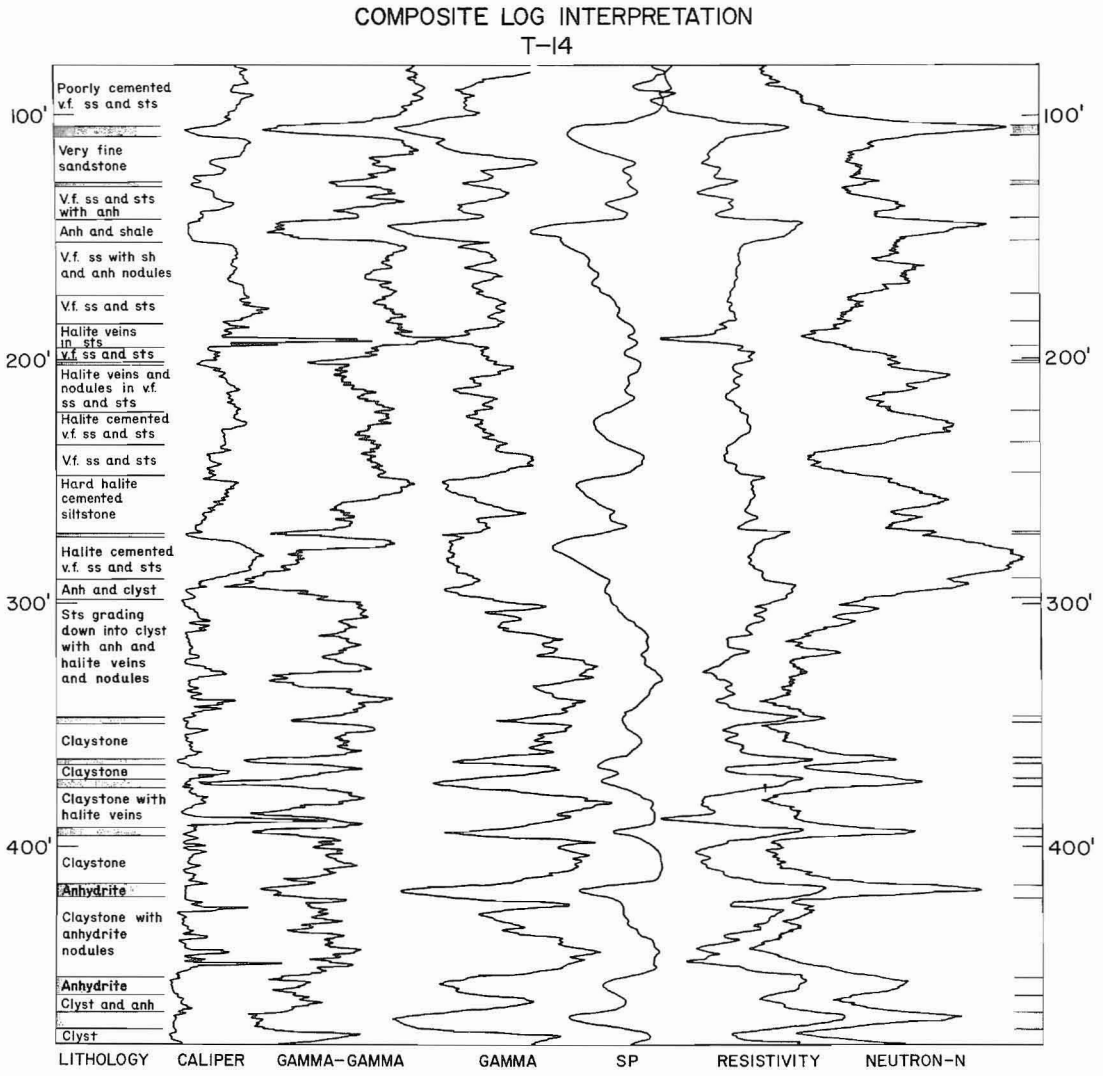


Figure 3. The relationship of log response to lithology and the similarities and differences between several different kinds of logs.

benefits expected from logging. In cased holes only the radiation logs can provide information on the groundwater system; however, each kind of radiation log measures a different parameter and when they are all used together one tends to support or contradict conclusions drawn from another.

Figure 3 shows six different logs run in a rotary test hole adjacent to a core hole on the upper Brazos River in Texas. The lithologic descriptions on the left were derived from the core. Note how the partially dissolved halite veins at a depth of 190 feet are reflected on the caliper, gamma-gamma and single-point resistivity logs. The significant increase in average porosity shown by the shift to the left on the neutron log below 300 feet does not reflect an increase in effective porosity or permeability because the gamma log shows this is merely an increase in clay content. A laboratory study of particle size distribution gave an average of less than 10 percent clay size (<0.004 mm) above 300 feet. Below 300 feet the average amount of clay-size particles is greater than 20 percent and two samples exceeded 50 percent. From these logs the zone above 100 feet was selected as having the highest permeability in the section illustrated and this was borne out by packer tests and pump tests. The zone above 100 feet shows the highest porosity on the neutron log, lowest resistivity on the single-point log, and the lowest clay content indicated by the natural gamma log.

Correction for shaliness or clay content is a very important use of the natural gamma log. The neutron log measures total porosity in saturated rocks, which in some areas is related to permeability. Shales and other fine-grained sedimentary rocks containing a high clay fraction tend to have a high total porosity but a low effective porosity and a low permeability. The gamma log usually indicates these beds and a correction factor for shaliness may be derived from experience in a given geologic environment. Bredehoeft (1964) found a consistent relationship between permeability of the Tensleep Sandstone and porosity derived from core analyses and neutron and sonic logs.

Figure 3 also illustrates the similarity of several pairs of logs made by entirely different sensing techniques; the resistivity log and neutron log and the gamma log and spontaneous potential log are very similar. If the gamma-gamma log were reversed, as is

common in petroleum logging, it would also be quite similar to the neutron log but with more hole-diameter effect. These similarities in response are quite useful in areas where radiation logs of cased wells offer the only means of correlating with preexisting electric logs.

A very important new technique of composite log interpretation has been pioneered by Schlumberger (Tittman, *et al.*, 1965). Charts are available which present a graphical comparison of neutron-derived porosities and gamma-gamma derived bulk densities or transit time from a sonic log. Porosities derived from these charts are corrected for the effect of matrix lithology. Furthermore, it is possible to identify the matrix mineral or the approximate percentage of two minerals in mixed lithology when two or more of these calibrated, porosity-sensitive logs are compared.

No sample log or suite of logs will provide the answers in every geohydrologic environment; therefore selection of the suite of logs most likely to provide the desired information is important. General guides to selection of a logging program are available for the various petroleum environments but experience in groundwater environments is still too limited for a categorical approach. The accompanying table is a list of required information and the logging techniques that might provide this information. Experience with logging and some knowledge of the geological framework, interstitial fluids and hole conditions are prerequisites to selection of the tools most likely to be productive. Many failures and negative attitudes toward logging are due to lack of recognition of these factors. Adequate logging equipment is important but will not produce results unless experienced, trained personnel are available throughout the program for planning, logging and log interpretation. The planning of a logging program should be based not only on a knowledge of the potential applications and limitations of the various devices but also on a thorough understanding of the information needed and the accuracy required to make this information useful.

The table can only serve as a quick reference guide; textbooks or reports on logging should be also consulted. The reference handbook of selected formation evaluation tools (SPWLA, 1966) is an excellent guide to commercially available logging services.

Suggested logging techniques for groundwater investigations.

Information Needed on Properties of Rocks, Fluid, Wells, or Groundwater System.	Conventional Logs which Might be Utilized.
Lithology of aquifers and associated rocks.	Electric, sonic, or caliper logs in open holes. Radiation logs in open or cased holes.
Stratigraphic correlation of aquifers and associated rocks.	Electric, sonic, or caliper logs in open holes. Radiation logs in open or cased holes.
Total porosity or bulk density.	Calibrated sonic or gamma-gamma logs in open holes. Calibrated neutron logs in open or cased holes.
Effective porosity or true resistivity.	Calibrated long normal or lateral resistivity logs.
Clay or shale content.	Natural-gamma logs.
Permeability.	No direct measurement by logging. May be related to porosity, injectivity, sonic amplitude.
Secondary permeability—location of fractures and solution openings.	Single-point resistivity, or caliper logs, sonic amplitude, borehole television.
Specific yield of unconfined aquifers.	Neutron logs calibrated in percent moisture.
Grain size.	Possible relationship to formation factor derived from electric logs. Clay content from gamma logs.
Location of water level or perched water outside of casing.	Electric, fluid resistivity, gamma-gamma logs in open hole or inside casing. Neutron or gamma-gamma logs outside casing.
Moisture content above water-table.	Neutron logs calibrated in percent moisture.
Rate of moisture infiltration.	Time interval neutron logs or radioactive tracers. Temperature.
Direction, velocity and path of groundwater flow.	Single well tracer techniques – point dilution and single well pulse. Multiwell tracer techniques.
Dispersion, dilution and movement of waste.	Fluid resistivity and temperature logs, gamma logs for radioactive wastes, sampler.
Source and movement of water in a well.	Injectivity profile, flowmeter, or tracer during pumping or injection. Differential temperature logs. Time interval, neutron, or gamma-gamma logs.
Chemical and physical characteristics of water—including salinity, temperature, density and viscosity.	Calibrated fluid resistivity and temperature logs in hole, neutron chloride logging outside casing.
Determining construction of existing wells, diameter and position of casing, perforations, screens.	Gamma-gamma and caliper logs, collar and perforation locator, borehole television.
Determining optimum length and setting for screen.	All logs providing data on lithology, water-bearing characteristics and correlation and thickness of aquifers.
Guide to cementing procedure and determining position of cement.	Caliper, temperature, or gamma-gamma logs.
Locating corroded casing.	Under some conditions caliper, or collar locator.
Locating casing leaks or plugged screen.	Tracers and flowmeter.

Logging techniques used in groundwater investigations

Electric logging. Electric logging was introduced in France by the Schlumberger brothers in 1927 and was the first geophysical well-logging technique. Electric logging includes numerous types of devices for simultaneously recording rock resistivity and natural electric potentials. Electric logs are run in almost every oil well. However, sophisticated electric logging techniques have not yet been widely applied to groundwater hydrology. Certain types of electric logs provide an estimate of effective porosity and water quality in granular aquifers and are excellent for determining lithology and stratigraphic correlation.

Spontaneous potential. The spontaneous potential curve is a record of natural voltages developed in most drilled wells between dissimilar fluids contained in the rocks penetrated and the borehole. The equipment is very simple; it consists of two lead electrodes, one moving in the drill hole and the other stationary at the surface. The recorder plots millivolt changes in electrical potential between these two electrodes as a function of depth.

The source of spontaneous or self potential (SP) in a drill hole has been the subject of much controversy but it is generally accepted to be the sum of electrochemical and electrokinetic or streaming potentials. Electrochemical potentials are due to differences in concentrations or activities of formation fluids and fluids in the hole.

The spontaneous potential curve may be used to calculate formation water resistivity, locate bed boundaries, distinguish between shales and sandstone or limestone in combination with other logs, and for stratigraphic correlation. The SP log is affected by hole diameter, bed thickness, water or mud resistivity, density, and chemical composition, mud cake thickness, mud filtrate invasion and well temperature. Although correction factors and curves are available to reduce or eliminate these effects considerable information obviously must be available to make the necessary corrections. The SP log is rarely used quantitatively in groundwater hydrology, but it is widely run for qualitative lithological information. SP deflections are read from a shale baseline on the right to maximum negative deflections. The shale baseline is drawn through as many deflection minima as possible. A sand line may then be drawn through negative deflection maxima and if fluid salinity is constant these lines will be parallel to each other and the zero baseline. Lithological contacts are located at points of maximum SP slope rather than at half amplitude as on many other logs. In many areas the SP log resembles the natural gamma log (Figure 3) and this characteristic is useful for correlating radiation logs of cased holes with electric logs. An unusual SP response is shown in Figure 1 where noise on the log from 720 feet to 760 feet was apparently caused by water entering the hole in this zone and moving up the hole.

Resistivity. Theoretically the resistivity values recorded on a log are a measurement of the resistance of a cube of material measuring 1 meter along each edge, hence the units are ohms meter^2 per meter or simply ohm-meters. Since most rocks consist of nonconductive particles, the nature of the pore spaces and interstitial fluids determines the character of the resistivity curve. The numerous types of resistivity curves made by commercial logging companies are differentiated by the configuration of the electrodes and the resulting differences in the thickness of rock units measured and the depth of investigation. The single-point resistivity log, along with the SP, is the most widely used logging techhique in water wells. Most single-point logging systems send 60 cycle AC to a lead electrode in the hole and the voltage drop or resistance between the electrode and body of the tool or cable sheath is measured. The single-point log is highly susceptible to borehole effects and has a very shallow radius of investigation. For these reasons it is very rarely interpreted quantitatively. The single-point resistivity log has several advantages over multiple electrode curves: it has a unique response, always giving a deflection to the right in beds of higher resistivity; it detects very thin beds and fracture zones and it has a logarithmic or nonlinear type response so that off-scale deflections are reduced.

Figure 1 illustrates the use of the single-point resistivity log to locate water-bearing fracture zones in crystalline rocks. Fractures are shown by sharp decreases in resistivity. This log shows fractures which were not detected by the caliper log and also suggests that some of the fault and fracture zones, located on the basis of a core log, have significant depth errors.

Multielectrode resistivity equipment is just beginning to be used on water well logging equipment, although logs run for petroleum are widely used for determination of groundwater quality by hydrologists in the southeastern United States. Short and long normal resistivity equipment is now available for single conductor cable equipment and changes to other electrode configurations can be made. Multiple electrode devices employ separate current and pick-up electrodes and spacing up to about 30 meters long is used. Current sent into the formation is usually 400 cycle AC and different sets of pick-up electrodes measure the resistivity of different vertical and annular volumes of rock. Radius of investigation varies from a few millimeters for a short-spaced contact device to many meters for a long-spaced device and this feature enables the determination of the extent of mud invasion which is related to permeability.

Commercial long-spaced resistivity curves are used routinely to determine the quality of groundwater according to the relationship $R_w = R_o/F$ (Turcan, 1962) where R_o is the resistivity of a rock saturated with conducting fluid, R_w is the resistivity of the saturating fluid and F is the formation resistivity factor which depends on the lithology of the aquifer. F is usually consistent for given sedimentary unit within a depositional basin and may be determined by laboratory analysis of samples or from the equation, when both resistivity logs and formation water samples are available. Effective porosity (ϕ) may also be determined from electric logs according to the following empirically derived equations: $F = R_o/R_w = 0.62/\phi^{2.15}$ for granular rocks. $\phi^{2.15}$ is porosity with a cementation factor which may vary from 1.3 to 2.2. For limestone and dolomite the relationship $F = 1/\phi^2$ is used.

Radiation logging

Natural gamma. Natural-gamma logs are continuous records of gamma radiation produced by radioisotopes naturally occurring in all rocks. Where borehole temperatures are less than 150°F gamma logs usually are made with a thallium activated sodium iodide detector; at higher temperatures the less sensitive Geiger tubes must be used. The gamma-emitting radioisotopes normally present are potassium 40 and the daughter products of uranium and thorium. Potassium 40 is probably the most common radioisotope in sedimentary rocks and the ratio of potassium 40 to stable potassium is usually constant. The total potassium content of rocks is extremely variable but the world-wide average for argillaceous sediments is about 3 percent and for sandstones is about 1 percent. For this reason it is generally true that the finer-grained detrital sediments in a section exhibit the maximum radiation on a gamma log but exceptions are common. The identification of shaley or clay-rich sediments with the gamma log is important in distinguishing sediments with a high total porosity from those with a high effective porosity. Evaporates and chemical sediments are generally lower in gamma activity than detrital sediments and mafic igneous rocks are usually lower than the felsic rocks.

The chief uses of the natural-gamma log in groundwater hydrology are stratigraphic correlation and identification of lithological units. It is the simplest and least expensive of the radiation logs and permits identification of aquifers in either cased or open holes. The gamma sonde is also used for location of radioactive wastes disposed of in the ground, in radioactive tracer work and in the evaluation of deposits of radioactive minerals.

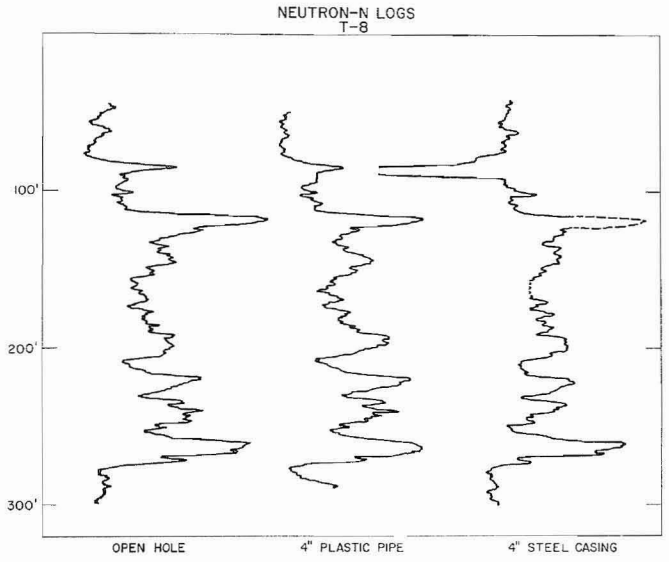


Figure 4. Three neutron logs made in the same hole under different conditions show essentially the same character. The obvious difference at 90 feet was apparently caused by removal of a thin anhydrite bed when the plastic pipe was being drilled out.

Neutron. The various types of neutron logs include potentially the most useful techniques in borehole geophysics as applied to groundwater investigations. This is because most of the log response is caused by changes in hydrogen content in the environment and in water wells this is primarily due to changes in water content. The neutron log is made with a neutron source attached to a probe containing either a neutron or a gamma detector. The neutron detector used in our research is a europium-activated lithium iodide crystal enriched in lithium 6. The gamma detector is a sodium iodide phosphor. Energy discrimination and cadmium and boron shields are used to make either neutron-epithermal neutron logs, in which most of the neutrons detected have an energy between 0.1 and 100 electron volts, or neutron-gamma logs where most of the system response is due to gamma photons produced by neutron capture or inelastic scattering. In early work on this project neutrons were derived from a 2-curie plutonium 239 - beryllium source; now a 3-curie americium 241 - beryllium source with an emission of 8.62×10^6 neutrons per second is used. These sources emit fast neutrons which are moderated (slowed) in the fluid column, casing and rock by collisions with atomic nuclei. The element most effective in moderating neutrons is hydrogen because the nucleus of a hydrogen atom has about the same mass as a neutron. The average energy lost by a neutron in a collision with a hydrogen nucleus is 63 percent as compared to 12 percent for the nucleus of an oxygen atom. Thus the range of a neutron is largely a function of the hydrogen content of the material through which it passes. In the long-spaced neutron probe used in well logging the number of neutrons detected and recorded is reduced by increasing hydrogen content of the environment.

Above the zone of saturation the calibrated and corrected neutron log provides a continuous measurement of the moisture content; in the zone of saturation it provides a record of porosity. An important new use of neutron moisture logs is the determination of the specific yield of an unconfined aquifer as described by Meyer (1963). The neutron-epithermal neutron logs are less affected by chemical changes and borehole diameter changes than the neutron-gamma logs. Figure 4 shows a series of three neutron-neutron logs of the same hole under three different hole conditions. With one exception the differences in the logs are minor. The plastic pipe had to be drilled out and some rock was also removed during this operation. The anhydrite bed shown at a depth of 90 feet on the logs made in the open hole and through plastic pipe was apparently partially removed during this operation and the third log indicates a large caved zone at this point.

The neutron-gamma probe has been found experimentally to be quite sensitive to chlorine content and so a comparison of neutron-epithermal-neutron logs, which are relatively insensitive to chemical composition of fluids, and the neutron-gamma log gives a theoretical basis for detecting major changes of chlorine content *in situ*, through casing. This technique has been used with partial success on the upper Brazos River in Texas to locate the interface between overlying fresh water and brine through casing. Early papers on the subject indicated that there was no chlorine effect on the gamma-gamma log. Theoretically it is true that there is no direct effect; however a brine containing as much as 200,000 parts per million total dissolved solids will be logged as a lower porosity, because of substitution of chlorine atoms for hydrogen atoms.

Gamma-gamma. Gamma radiation is attenuated or scattered by all material through which it travels as a function of the electron density of the material. The gamma-gamma log is a record of changes in detected gamma radiation, derived from a source in the probe, caused by differences in the density of material in and adjacent to the hole. In logging equipment used by the Water Resources Division 10 to 35 millicuries of $cobalt^{60}$ is used as a gamma source attached below a sodium iodide detector. Both an axially symmetric and a decentralized side collimated tool are used. The axially symmetric tool employs Mallory 1000 metal shielding between the source and detector and is not positioned in the hole. The decentralized device employs two motor-driven arms to force the tool against the wall of the hole. Mallory 1000 metal is used as a shield around the source and detector so that most of the signal recorded is from the portion of the hole wall in contact with the tool.

The decentralized tool has the advantage of being much less affected by changes in drill bit size or position in the hole. It is, however, still affected by hole diameter changes occurring between the source and detector. Figure 5 illustrates the effect of hole diameter on gamma-gamma logs made with the axially symmetric tool. The logs plotted with a solid line were made in a core hole where excessive circulation caused the caving observed. The dashed logs were made in a nearby hole drilled with a tricone rotary bit and a minimum of circulation.

Gamma photons from the source penetrate, are scattered and absorbed by fluid in the hole, casing and formation. The relative percentage of photons absorbed and scattered and radius of investigation depends greatly on the type and size of the source, spacing between the source and detector, distance to the wall of the hole and material penetrated. With the equipment and source described, Compton scattering is probably the most important

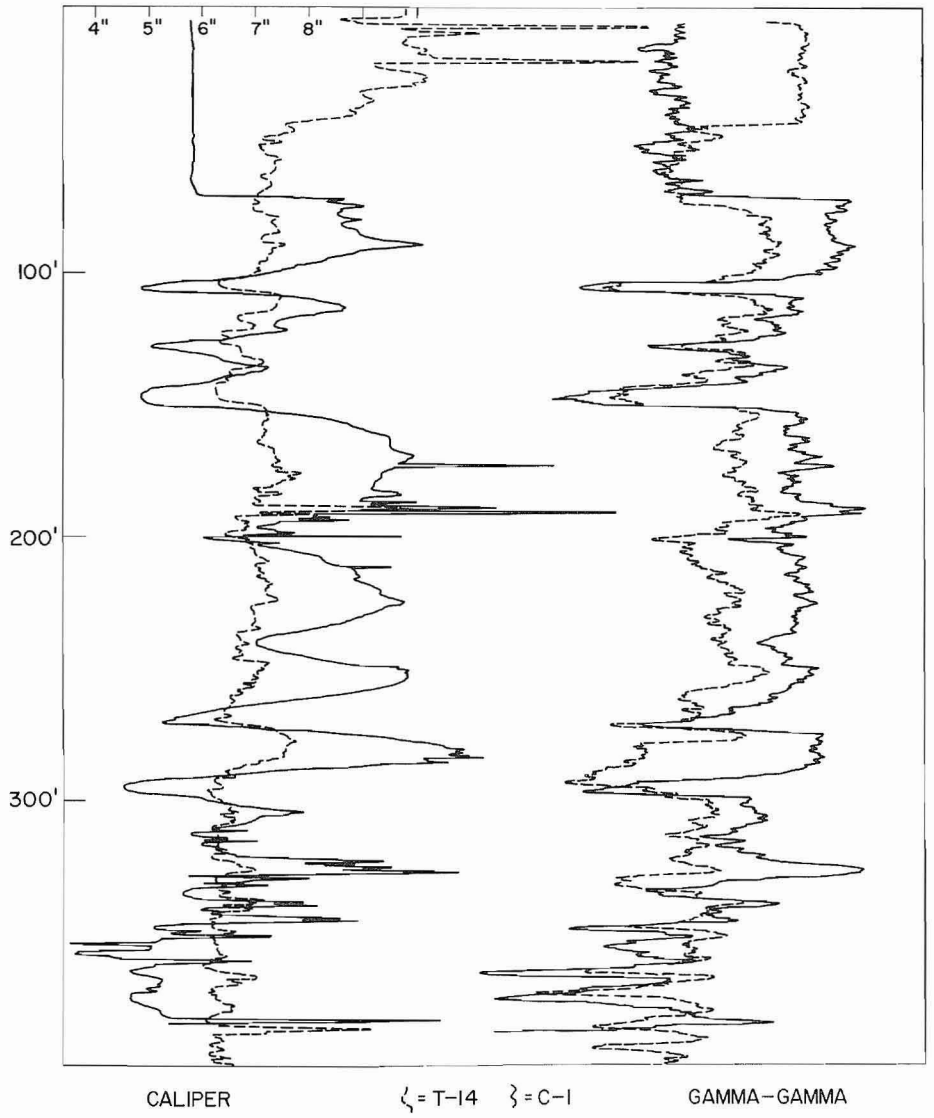


Figure 5. Caliper logs in adjacent holes with the same lithology demonstrate the effect of different drilling techniques on the character of the holes. Gamma-gamma logs of these same two holes illustrate the response caused by these changes.

process of photon energy degradation taking place. Therefore gamma radiation absorbed is proportional to the electron density of the material penetrated and the electron density is approximately proportional to the bulk density of most material encountered in logging. The count rate plotted on a gamma-gamma log is an exponential function of bulk density. The calibrated and corrected gamma-gamma log may be used for the determination of porosity by the application of the following equation:

$$\text{Porosity} = \frac{\text{grain density} - \text{bulk density (from log)}}{\text{grain density} - \text{fluid density}}$$

Grain density may be derived from laboratory analysis of samples or, in the case of quartz sandstone, a value of 2.6 gm/cc may be assumed. The fluid density in most water wells may be assumed to be 1 gm/cc. If a high content of dissolved solids is suspected, a laboratory analysis may be necessary. Errors in bulk density obtained by the gamma-gamma method may be as high as ± 0.04 gm/cc. A linear plot is obtained when bulk density is plotted against the logarithm of count rate.

In addition to measurement of bulk density or calculation of porosity the gamma-gamma log may be used to locate casing or collars, including one string of casing outside another. It may also be used to locate the position of grout or hole enlargements through casing and detect the water level and significant changes in fluid density.

Caliper log. The caliper or hole-diameter log is essential to the correct interpretation of other logs and has considerable use in providing information on hole construction and lithology. Because most lithological geophysical logs are affected to some degree by hole diameter, the caliper should be included in any open-hole logging program. The caliper used on this research project has three motor-driven arms that may be opened or closed from the surface. The tool is lowered to the bottom of the hole, the arms are opened by a motor and maintained against the wall of the hole by spring pressure. A continuous record of average hole diameter is made as the tool is raised. Accuracy and reproducibility is better than .25 inch of hole diameter and three different lengths of interchangeable arms permit a high degree of sensitivity in holes from 2 to 36 inches in diameter. With the caliper equipment described it has been possible to detect casing collars, very small fractures and vesicular tops of basalt flows.

Bow spring type calipers are much less sensitive and record only major changes in hole diameter.

The caliper log is a direct measurement of average hole diameter and is not subject to extraneous effects except heavy mud which may prevent the arms from contacting the walls of the hole. Mud-cake buildup detected by the caliper may be interpreted in terms of relative permeability. Hole-diameter changes are a function of the drilling technique: speed of rotation, pressure on the bit, circulation pressure and rate, etc., and lithology; hardness, compaction and cementing. The caliper logs shown on the left of Figure 5 illustrate both drilling and lithological factors. The differences in hole diameter are obvious but the similarities are also important. The two logs correlate very well, the thin washed-out zones just above 200 feet represent halite veins averaging 1 inch thick. This zone of veins probably represents an unconformity and has been correlated for a distance of several miles by the use of caliper logs. In other holes it is a source of brine and gas.

Sonic velocity. A system for running borehole-corrected continuous sonic-velocity logs on single conductor cable has recently been developed and is now being used experimentally on our research logger. Sonic-velocity logs are probably the most widely used in the petroleum industry for the measurement of porosity, and recently developed techniques for log collation offer the possibility of obtaining more accurate porosities and matrix lithology when sonic logs are compared with gamma-gamma and neutron logs. The sonic pulses are generated by magneto-restrictive transducers and have a frequency of about 20 kilocycles. In the tool used by the Water Resources Division two transmitters are located respectively 3 feet above and below two receivers which may be spaced either 1 or 2 feet apart. Two sets of bow springs are used to keep the sonde centralized in the hole. The log is an average of the transit time between the receivers, of an upward travelling pulse and a downward travelling pulse. The purpose of the dual transmitter-receiver system is to eliminate excursions to the right and left on the log caused by differences in the distances from the hole wall to the 2 receivers as the tool enters and leaves a cavity. Although new types of sonic logging systems are called borehole-corrected or compensated it is doubtful that diameter effects are completely eliminated; therefore a caliper log is a useful adjunct to interpretation.

The sonic-velocity log is recorded as transit time (Δt) in microseconds per foot, with zero on the right. Longer transit time (lower velocity) indicates a higher porosity, therefore the sonic log has the same general character as neutron, reversed gamma-gamma, and resistivity logs with porosity increasing to the left. Porosity is calculated from sonic logs using the following empirical equation:

$$\text{Porosity} = \frac{\Delta_t \log - \Delta_t \text{ matrix}}{\Delta_t \text{ liquid} - \Delta_t \text{ matrix}}$$

Tables are available giving average sonic velocities for liquids and various rocks. Water or mud has a velocity of 5000 to 5300 ft/sec (200 to 189μ sec/ft), shale from 6000 to 1600 ft/sec (167 to 62.5μ sec/ft) and dolomite about 24,000 ft/sec (42μ sec/ft). Graphs are also available from the logging service companies for direct reading of porosity when matrix lithology is known. Sonic-velocity data are of considerable value as an aid to the interpretation of surface seismic profiles.

A second general type of acoustic logging is the recording of signal attenuation or amplitude. Sonic-amplitude logging or full wave sonic recording is used in cased holes to determine the quality of the bond between casing and cement, and cement and hole wall. A relationship may also exist between formation character and sonic signal recorded through casing. Sonic attenuation is also an important means of locating zones of secondary porosity such as fractures and solution cavities in limestone. Various types of acoustic logging devices and log interpretation currently are the subject of much research and it is expected to become one of the most useful logging techniques in future groundwater investigations.

Fluid logging. Fluid logging includes the use of sondes to measure the temperature, quality and movement of fluids in a drill hole. These characteristics of the fluid column may or may not truly reflect conditions in the aquifer system. A considerable length of time is required for the water in most wells to reach thermal and chemical equilibrium with the surrounding rocks. Each time a probe or sampler is moved through the well these conditions are disturbed. For this reason the most valid temperature and quality logs are made on the first trip down the hole and the log made coming back up seldom repeats exactly. In addition vertical movement of water in the hole is commonly caused by differences in head among artesian aquifers. Under such conditions water in the hole may be representative of a single aquifer rather than rocks adjacent to the measurement and flowmeter or tracer logs are essential to the correct interpretation of fluid character logs.

Temperature. Our research logging equipment utilizes a single probe to make simultaneous records of fluid temperature and resistivity on the first trip down the hole. Temperature is measured with a calibrated thermistor inside a protective tube so designed that water is forced through the resistivity electrodes and past the thermistors as the probe is lowered. Calibration of the thermistor may be checked with a resistance decade box.

A differential temperature logging system has been used experimentally on a lease basis and results indicate that this may be a very useful logging technique in groundwater hydrology. A simultaneous gradient and differential temperature logging system for our research logging equipment is presently being tested. Differential logging systems are of two types. One system employs two thermistors spaced from one to several feet apart along the axis of the drill hole (Bashman and Macune, 1952). The differential curve is a record of changes in the temperature difference between these two thermistors. Another system employs one thermal sensor with a memory device to compare the temperature at a point in time with a preselected past time (Johns, 1966).

The first system is subject to error due to disturbance of thermal conditions by the lower thermistor. The second system is subject to error caused by changes in logging speed. Both systems have a theoretical sensitivity of .001°F, much more than conventional gradient logging systems. It is much easier to measure small differences in a factor on a differential than on an absolute scale. Gradient or absolute scales should be selected to cover the temperature ranges expected in the hole and thus reduce scale-changing to a minimum. The differential settings are selected so that a reference gradient near the top of the hole is recorded at a zero baseline and slight increases or decreases in

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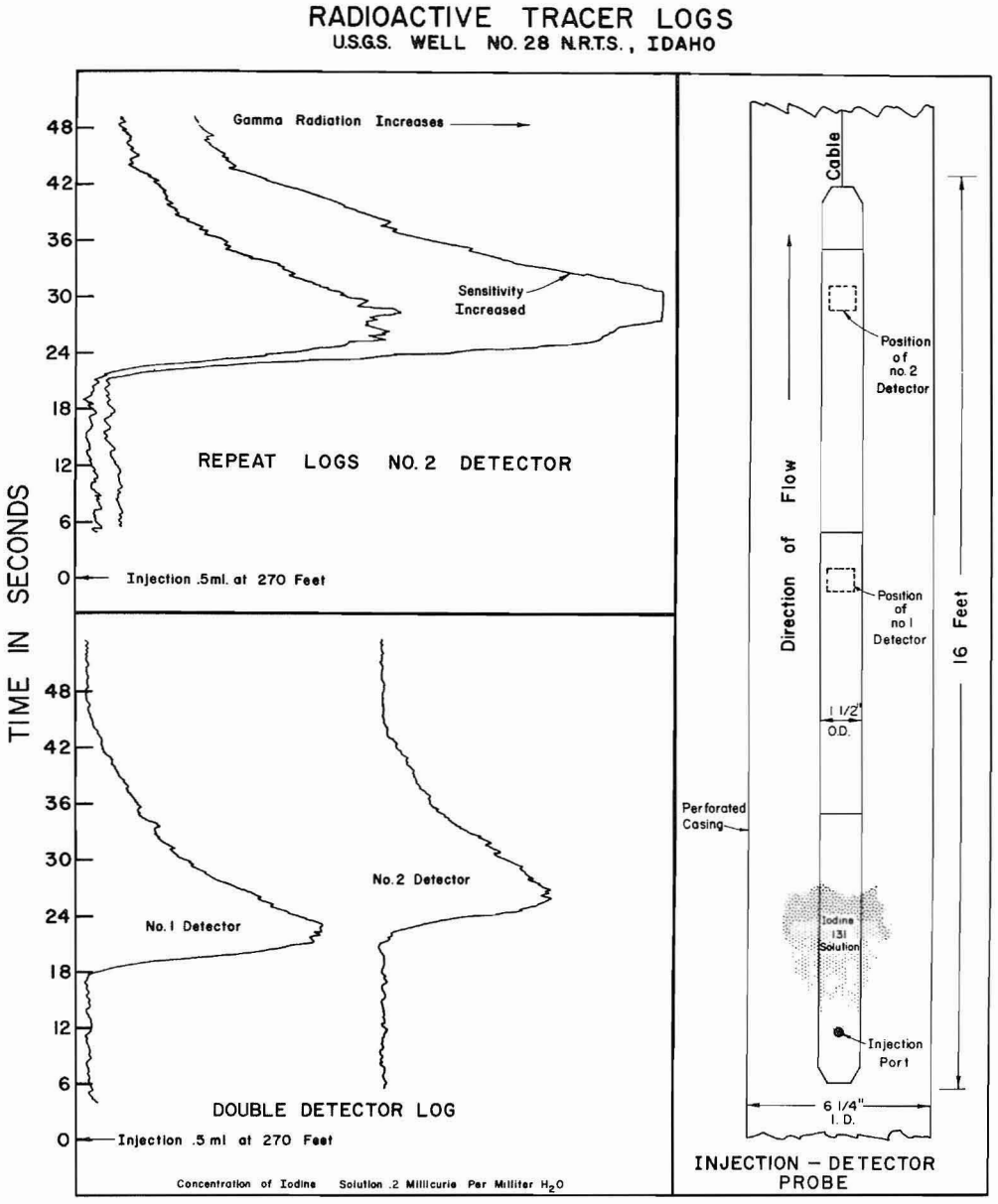


Figure 6. The double-detector radioactive tracer ejector tool is shown on the right. The two single-detector logs shown on the top left indicate the repeatability of velocities measured by this technique. The bottom left is an actual double-detector log made at the National Reactor Testing Station in Idaho.

gradient are recorded as sharp deflections to the right or left. Most of these changes are not detectable on the gradient log.

The geothermal gradient is recorded on a temperature log of a well that is in thermal equilibrium with the adjacent rocks where there is no vertical fluid circulation. This gradient is largely a function of the thermal conductivity of the rocks modified by water moving through the system. Temperature logs may be used to identify aquifers or perforated sections contributing water or gas to the well; to provide information on the source of water; as an aid in identifying rock types; and for calculating fluid viscosity and specific conductance from fluid-resistivity logs. Temperature logs made before and after pumping or injection may identify the aquifers and suggest the relative permeabilities. The sensitivity of the differential device suggests the use of water itself as a tracer. Temperature logs have been used to distinguish moving and stagnant water in a well and identify the source of recharge or injected waste water. Seasonal recharge may be reflected in cyclic temperature fluctuations and it is possible to interpret the vertical movement of water in the unsaturated zone from very sensitive temperature logs (Stallman, 1965).

Fluid resistivity. Fluid-resistivity logging is the measurement of resistivity, in ohm-meters, of fluid between two noble-metal electrodes in the hole. These values may be converted to specific conductance if temperature information is also available, or to total dissolved solids if ions present are identified. The fluid-resistivity log may be used to locate points of influx or egress of waters of differing quality, to locate the interface between salt water and fresh water in a well, correct head measurements for fluid density differences, locate waste waters and follow the movement of saline tracers. The resistivity of the fluid column is also important in interpreting SP, resistivity, and neutron logs which may be effected by salinity changes.

Flowmeter and tracer logging. Devices used on this project to measure vertical flow in water wells include the impeller flowmeter, the radioactive tracer ejector-detector and the brine ejector-detector. Vertical flow measurements with these devices have been used to measure relative aquifer permeabilities under pumping or injecting conditions. They are also used to establish the location of leaks in casing and as an aid in interpretation of other logs.

The impeller flowmeter is the simplest and least expensive device but it is limited to higher velocities and must be carefully calibrated for quantitative use. It may be used for a continuous qualitative log if tool movement can be maintained fairly constant. In this mode a comparison of down and up logs of integrated revolutions per unit time will locate zones of flow and indicate the direction. Revolutions recorded on time drive with the flowmeter held stationary in the hole will then provide more accurate measurement of flow at points of interest. The flowmeter must be calibrated for the hole or casing size and errors are introduced by sonde position in the hole.

The radioactive tracer ejector-detector provides a system for measuring in-hole velocities of a few feet per day (Keys, 1966). It is also the only means of measuring movement behind casing. The tracer ejector consists of a positive displacement piston-type pump that will eject from a fraction of a milliliter to 200 milliliters of water-soluble tracer (Figure 6). An aqueous solution of iodine 131 is used for most of this work. The same radiation probes used for natural-gamma logs may be attached to the ejector in any one of a number of combinations. When direction of flow is not known they may be placed above and below the ejector. When direction is known they may both be attached above or below the ejector to increase the accuracy of measurement as shown at the lower left of Figure 6. With this system vertical flows as high as 30 feet per minute have been measured in wells in basalt at the national Reactor Testing Station in Idaho and horizontal flow in some of these aquifers removes all of the tracer from the well in several seconds. In contrast, vertical velocities of a few feet per day were measured in fractured crystalline rocks at the A.E.C. Savannah River Plant in South Carolina.

A brine ejector-detector system utilizing the same ejector tool and one or several fluid resistivity tools as detectors has been recently developed. It has the advantage of utilizing a less expensive, more readily available tracer that does not require licensing. However, it has disadvantages in being more affected by the specific gravity of the tracer, a lower detection sensitivity and not being usable for detecting flow behind casing. Recording is done in exactly the same way as for the radioactive tracer and the logs look nearly the same. An event marker records the instant of ejection and a plot of resistivity versus time is made automatically.

A number of other in-hole tracer techniques have been used to measure relative permeability. These include insoluble radioactive tracers which are concentrated in the most permeable beds under fluid injection conditions and a single-well pulse technique relating the recovery time under pumping conditions of a slug of tracer placed in the wall under injection conditions (Borowczyk, *et al.*, 1967).

Collar and perforation locator. The magnetic collar locator is a device that may be run simultaneously with radiation logs to accurately relate lithological changes to the depth of casing collars. This use is of particular importance in deep holes where cable stretch may introduce logging footage errors. However it depends on accurate measurement of casing by the driller. The sonde contains a permanent magnet inside a coil. As the probe is moved in the hole, steel casing cutting the magnetic flux lines induces a small voltage in the coil which is recorded at the surface. Changes in the mass of metal within the field, such as collars or perforations, cause differences in the potential measured. An event marker may be preset to record a kick representing a collar above a certain millivolt threshold or a continuous record may be made so that collars or perforators may be picked at a later time. This tool is most useful in logging of old water wells where accurate information on the position of casing, collars and perforations is not available and is needed to aid in the correct interpretation of other logs.

Logging equipment

A most important phase of any research program on the application of logging to groundwater is development of lightweight, inexpensive equipment. Such equipment is only feasible because most water wells are not as deep as oil wells and temperatures and pressures are correspondingly lower. Solid state electronics, from the transistor to microcircuits, have made it possible to install a very complex logging system in a suitcase and most probes can be made less than 1 inch outside diameter. Reduction in equipment size not only lowers packaging, transportation and power costs but makes it possible for one man to make logs that require two or three men in the petroleum industry. The equipment described can produce logs equal or superior in quality to those produced by petroleum logging equipment with a significant reduction in cost.

All of the logging equipment owned by the Water Resources Division of the U.S. Geological Survey utilizes single-conductor, double-armored coaxial cable. This cable has several advantages. It is less expensive than multiple-conductor cable and much lighter so that winch size and power requirements are reduced. Logging devices developed for use on single-conductor cable may be adapted to any logger; this is not true of most multiple-conductor devices. A major objection to the use of single-conductor cable has been the need for much more complicated electronics in the sondes. This factor is becoming much less important with the rapid advances taking place in electronics and the availability of microcircuits. Another advantage of multiple-conductor cable formerly has been the ability to make a number of logs simultaneously, an important cost factor in deep oil wells. With recent advances in telemetry it is now possible to make a number of logs simultaneously with single-conductor cable.

Graphic recorders utilizing pen and ink are widely used in the small loggers instead of the multitrace photographic type recorders employed in petroleum logging equipment. Pen and ink recorders are cheaper, easier to maintain and more adaptable to the varying needs of the groundwater industry. Furthermore, they provide a record which is available for interpretation as it is being made.

The multifunction logging equipment used in the Water Resources Division research project was developed specifically to meet the project objectives as outlined above. It consists of a carryall-mounted 6000-foot logger and a suitcase-contained 500-foot minilogger. Specifications were prepared by project

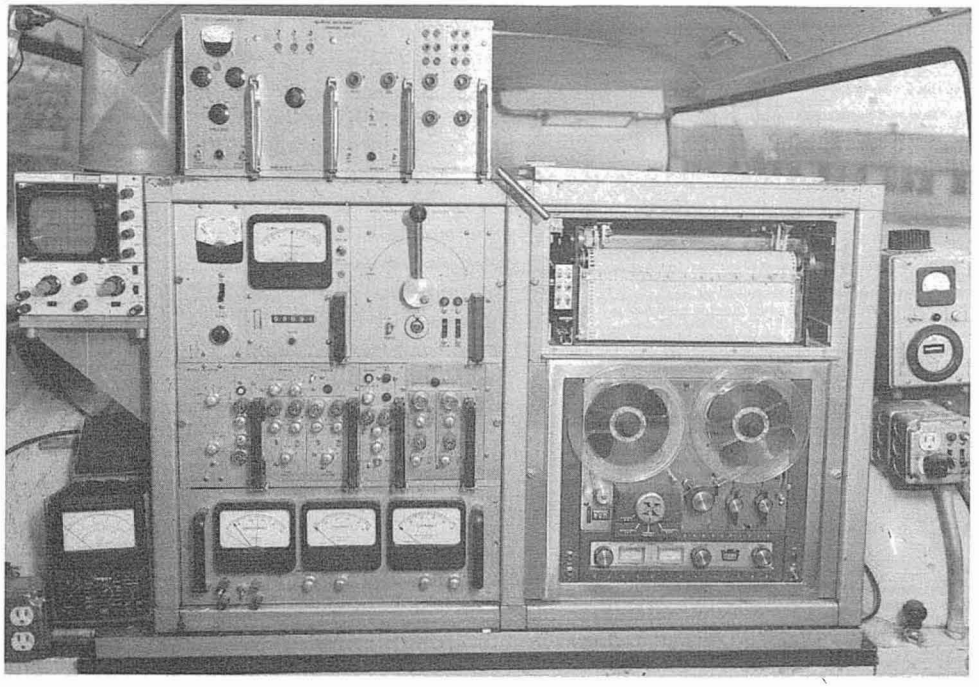


Figure 7. Control panel for vehicle-mounted logger. Scope for trouble shooting and sonic calibration is on the far left. Control modules for all types of logging and winch operation are on the left. Graphic and tape recorders are on the right and power frequency and voltage controls, far right.

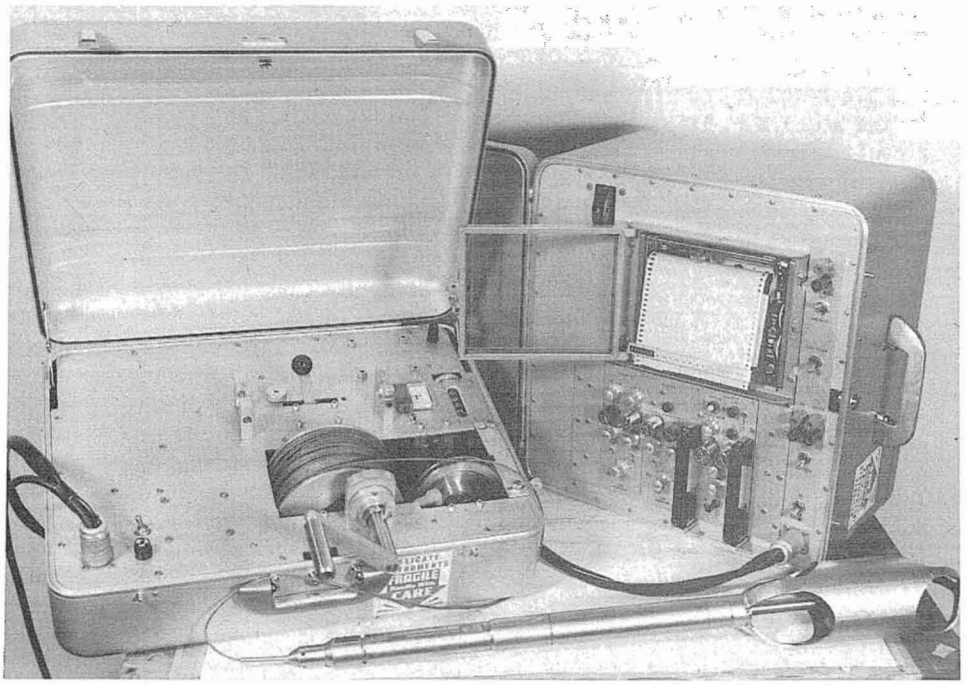


Figure 8. Suitcase-mounted minilogger with impeller flowmeter in the foreground. Plug-in control modules below the recorder on the right are also used in the vehicle-mounted logger.

personnel and the equipment was built by Well Reconnaissance, Inc. of Dallas, Texas. Many features were modified several times during construction and testing, and improvements are still being made by project personnel. The basic modular design was developed for ease of modification and addition of new logging functions (Figure 7). A minilogger, carried in two suitcases, utilizes the same modules and probes in a readily transportable unit that weighs less than 100 pounds (Figure 8).

The probes are constructed of stainless steel, double *O*-ring sealed, and pressured tested to 6000 pounds per square inch. Most of them are 1 1/2 inches in outside diameter, and they range from 2 to 6 feet in length. Downhole electronics are designed to operate at temperatures of 0 to 150° F. One probe emitting both positive and negative pulses is used for simultaneous fluid temperature and resistivity logs and as an attachment to the tracer ejector for brine tracer logs. Two electronic sections emitting positive and negative pulses are employed with three separate and interchangeable detector sections for radiation measurement. Scintillation crystals and coupled photomultiplier tubes serve as detectors in both gamma and neutron devices. The

sondes are inserted in portable field standards to record system drift. Two sleeve-type uranium sources are used as field gamma standards. The neutron field standard employs an air-filled aluminum cylinder which may be placed in several different positions inside a larger water-filled cylinder to simulate various porosities. Both these standards have been related to the American Petroleum Institute calibration pits at Houston, Texas.

A system for recording all radiation logs on magnetic tape was developed and built by E.R. Bullard as part of this project (Keys, 1967). An inexpensive, unmodified two-track stereo tape recorder is utilized. The raw data, as received from the probes, is recorded on one channel and a footage-synchronizing signal is recorded on the other channel. A standard log is plotted on the graphic recorder at the same time. Either positive or negative pulses are taken off at the top of the cable and recorded prior to signal modification in the control modules. Footage-synchronizing pulses consist of two 1-kc pulses per foot of hole logged.

Playback from tape to graphic recorder may be done in either the field or the office with either the truck-mounted logger or the minilogger. In the playback mode the raw data from the log channel is fed through the control module where the following adjustments may be made: vertical scale, horizontal scale or sensitivity, zero positioning, time constant and recorder gain. These are the same controls that are normally adjusted when logging a well.

The applications and advantages of direct tape recording of geophysical logs are numerous. It is an aid to log interpretation because changes in horizontal and vertical scale to conform to other logs are easily made in the office. Similar adjustments may be made to bring a log into register with calibration standards. Correction for common operator errors is possible. One of the most common of these is loss of record because of movement of the pen off scale in either direction. With a tape-recording system this record is available for replay at any time. Magnetic tape can be fed directly into an analog-to-digital converter as a first step in the computer analysis of logs. The system also provides economical storage of easily retrievable data. As many as 16 logs of wells averaging 500 feet deep have been recorded on only one side of a 5-inch reel of tape.

Recently developed techniques with potential application to groundwater

Several logging and interpretive techniques now being used experimentally in petroleum logging have considerable potential for groundwater investigations. These more sophisticated tools and others that will be developed in the future will be widely used in hydrology only when the equipment is perfected for routine use and size and cost are reduced. Currently the techniques with the greatest potential appear to be neutron life-time logging, several types of spectral logging, acoustic amplitude logging, nuclear magnetic logging, and computer analysis of logs.

Neutron life-time logging employs a pulsed neutron generator and a synchronously gated gamma or thermal neutron detector (Youmans, *et al.*, 1966). With this system the neutron or gamma field is measured at one or two preselected times after a pulse of neutrons is emitted by the generator. A consistent relationship has been demonstrated between the pulsed neutron log response and formation porosity and fluid saturation. The system has several advantages over the continuous neutron logs. It is much less affected by borehole parameters, produces more accurate quantitative results in cased holes and the generator is capable of higher neutron fluxes than a radioactive source yet requires no shielding when it is turned off.

A similar type of portable neutron generator makes it possible to do in-hole neutron activation analysis. Gamma rays with an energy characteristic of the element are emitted when neutrons are thermalized and captured. A multichannel analyzer provides a record of the energy distribution of the gamma photons that reach the scintillation detector. Although borehole effects introduce many complications this technique has been used for qualitative activation analysis in drill holes. Neutron generators now available are too large, expensive, and require too much electrical power for widespread use in hydrology.

A multichannel spectrum analyzer and energy-discriminating logging equipment also provide means for deriving more information from natural-gamma logs. Potassium 40 and the daughters of uranium and thorium present in different quantities in different rocks and their relative abundances can be used to identify rock units. Spectral equipment has also been used to identify radioisotopes present in liquid wastes disposed to the ground and radioactive tracers in the groundwater system. Spectral analysis and energy discrimination offer means of improving the quality of neutron and gamma-gamma logs. Threshold and upper level discriminators can be adjusted so that most radiation below or above these energy levels is rejected.

Acoustic amplitude logs are a record of changes in the amplitude of the received sonic signals instead of the travel time of a pulse as in sonic velocity logging. Attenuation of sonic amplitude is used to interpret the quality of bonding between cement in the annulus and the casing and rock and to locate zones of secondary porosity such as solution openings or fractures. The relationship of sonic attenuation and the lithological characteristics of aquifers which effect permeability should be investigated.

Nuclear magnetic logging is a relatively new technique that has not yet been applied to groundwater investigations and the logs are available only from a commercial logging company. A relationship between parameters derived from nuclear magnetic measurements and the permeability of a number of sandstone samples has been recently demonstrated (Seevers, 1966). The logging device measures the nuclear magnetic properties of hydrogen in formation fluids and provides two values: the free-fluid index which is related to effective porosity and also to the amount of fluid free to move, and thermal relaxation time. Both the theory and equipment of nuclear magnetic logging are too complex to be adequately described here. Briefly, the down-hole equipment consists of a coil powered from the surface which creates a strong magnetic field, and switching and amplifying circuits. The magnetic field causes some spinning nuclei of hydrogen atoms to realign themselves with the artificial field. Precession of the spinning protons when the field is cut off induces a weak AC voltage in the coils of the inactive electromagnet. The most important factor in this logging system is that it is almost uniquely responsive to hydrogen contained in low-viscosity fluids. Furthermore, this does not include a detectable response from fluids mechanically bound in fine-grained detrital sediments and not capable of moving. This is in contrast

to the conventional neutron log which responds to hydrogen regardless of the chemical or physical form in which it occurs.

Computer interpretation and collation of geophysical logs is now available from commercial well logging companies and computer interpretation of logs is also used in uranium ore reserve calculations by the U.S. Atomic Energy Commission (Dodd, 1966). Some factors such as corrected total porosity, effective porosity, grain density, shaliness and true resistivity may be calculated from logs digitally recorded on magnetic tape. These factors may be played back as a continuous record and the mode of graphical presentation can be selected to fit the problem. Existing geophysical logs can also be digitized and recorded on magnetic tape so they are amenable to computer analysis. To justify programming and computer time for the interpretation and collation of groundwater logs a large volume of standardized logs must be available. The few experimental logs now available probably do not warrant such an expense. However, one use of computers that may be justified now is the calculation of correlation coefficients as a means of testing the depth fit of core or sample values to geophysical logs (Jeffries, 1966).

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Discussion on Borehole Geophysics as Applied to Groundwater, by W. Scott Keys

Dyck, John, Saskatchewan Research Council, Canada. We have used well logging in Saskatchewan for particular evaluation of formations in boreholes put down on test drilling programs some time ago. We have been able to establish some very valuable parameters. I have noted off the cuff remarks about the small amount of money that has been invested in the water well industry. I would like to ask Mr. Keys and people from other countries just how much money is invested in the water well industry, first of all, in the United States and anywhere else?

Keys, W.S. I will use a figure that was published by the Water Well Users' Association. In 1964, there were 464,000 water wells drilled in the U.S. by about 13,000 water well contractors, separate contractors not drill rigs, and the total investment in the wells not including pumps and plugins is over $500,000 in that one year. I suspect that at the present time, we are talking about over 500,000 wells per year at somewhat higher cost. The average cost of a well in the U.S. is $1500, so you cannot invest too much in logging wells of this type unless you have a specific problem to solve. You cannot interest the property owner who has spent $1500 for a well in a $1500 geophysical well-logging program. But as the wells get deeper, as they are doing, we had in the western U.S. municipal water supply wells that were 4000 feet deep. These wells were comparable in cost to oil wells and they are at the present time justifying whether expensive logging programs can be brought in for guiding the construction of these water wells.

Hurr, R.T., U.S. Geological Survey. I don't imagine these municipalities would be interested in logging their wells if they did not anticipate at a future time that they would probably have to put in more wells due to increase in population. Do you suppose that this is part of their stimulus?

Keys. I wish I could say that they were thinking that far into the future. I think that most of them are a little more short-sighted. A lot of these deep wells are in areas of oil wells and a lot of these people and drillers know something about geophysical logging. The logs are largely used to locate the aquifer, give information on its thickness and water quality, and possibly they may use the technique of casing the well and jet perforate it at the zone selecting by electric and gamma logging. I don't know of any municipality getting information for use in the future.

Crosby, J., Washington State University, U.S.A. I feel I ought to say a few things here regarding some of the concepts that have been brought up. To Mr. Meneley, I want to say first of all, I wouldn't trade a trunk load of seismic data for one good drill hole. I think maybe we are losing sight of the cost factor here. After listening to Mr. Meneley say that he can drill from $1 to $2 per foot, I can see why he wants to do all his work by drilling. Back in Washington State, the farmer goes out and puts a drill hole in the basalts for groundwater and he ends up with a bill from $2000 to $2500 just for putting in a hole. On the other hand, we just let a contract to put down a municipal hole for $80,000 for drilling only. We are up there in the same category as oil well drilling and I think you have to form an entirely different framework to your thinking.

We have used geophysics quite extensively and are quite well satisfied with it. We are certain that we shall have no trouble whatsoever in selling the communities on logging wells as a basis for helping their future drilling, not when you've got $80 to $100,000 tied up in a hole. Out in the central part of the plateau where the basalts are thick, we know the basic hydrological factors and once we reach the water-table level we know that we can probably get a gallon per minute per foot of penetration. But when we get over on the fringes of the plateau where we run into the old land surface, we have to be very accurate in putting our holes down in the old buried valleys. We have to pick these up with the use of geophysics. All the geologic mapping in the world will not locate them. It is a matter of cost more than anything else.

Keys. Well, I would stress cost in logging because the logging equipment has had to be developed in our business. Commercial logging trucks make the cost incompatible for use in water wells. We have to consider the economics of these techniques.

Shiftan, Z., Tahal Consulting Engineers, Israel. You speak about logging, in this case of cost.

Keys. Yes, geophysical well logging using existing commercial techniques. The costs applied to oil well logging are too high for most water wells. Some water wells are now logged by commercial oil well units. For the average project, such as for the USGS, we cannot afford the high costs. Most of the commercial tools are 3 5/8 inches in diameter while the diameter of many of our test holes is smaller. The time and transportation factors of commercial loggers must also be taken into account. Also they are available only in oil areas. Therefore, there are a number of reasons why we are not able to use commercial equipment for water wells. There are, however, in the United States a number of small companies making loggers that can be used in water wells.

Shiftan. I am sorry, but I never met the argument that logging of wells is too expensive. For the shallow wells, we always use the Widco logger which we operate ourselves. For the deeper wells, we always employ service companies who are normally in the oil business and I have never heard of any objections as to cost.

Keys. You are talking about large government programs. The U.S. Government does not drill water wells except on government reservations and AEC property, etc. I am talking about wells drilled by small towns and independent property owners. They cannot afford a logging consultant's help.

Collett, L.S., Geological Survey of Canada. I wonder if we should define what we mean by logging. Is it SP and single-point electrode that you were speaking of, Mr. Shiftan?

Shiftan. For the not too deep wells, certainly. But for the deeper wells, those going down to 3000 feet, we did all the logs that we could think of which would be appropriate in that case.

Keys. There is quite a difference in cost doing it commercially than doing your own. It is much less doing resistivity and SP, of course, than running the whole sweep. We, in our research, run the whole sweep of logs. There are some deep wells where we use commercial logs. For example, at Norfolk, Virginia, where a hole is put down through the coastal plain sediments, Schlumberger are going to run some logs and we are going to run some of our logs for comparison purposes. In this case, the expense of the well justifies the expense of the logging.

Hurr. Can you give any idea how many wells were logged by the USGS last year and how many were private versus government wells?

Keys. Most of the wells that we logged last year were private wells.

Hurr. Were they drilled for the purpose of the USGS?

Keys. They were definitely drilled for our purposes. I have a research project where I work all over the 50 states. We are trying to investigate the application of logging techniques to various groundwater environments that we have around the country. I only go out and log when it seems that I might learn something new about these various logging techniques. We have about 40 geophysical loggers available in the Water Resources Division. They do not do logging for private individuals who have problems. They only log wells when we need some objective information. I have very little use for drillers' logs which are more often than not misleading. Eventually at least, a single-point resistivity and SP will be required because we are losing a very important source of subsurface information that may be useful at some time in the future.

Jennings, C.M., Geological Survey of Botswana. Could you tell us the cost of the USGS logger that you have in your truck?

Keys. Not really because that unit has been modified so many times. The original logger cost only $4000. We put another $24,000 on it for additional tools. Unfortunately the companies that manufacture small loggers do not have too many on-the-shelf items. You have to tell them what you want in the way of logging gear. It used to be that you had to deal with one small company which was Widco. You could get a 3000-foot electric and gamma-ray logger for less than $10,000. Then on adding tools, each one will cost at least another $2000. This is cheap compared to the big commercial service company loggers.

Morley, L.W., Geological Survey of Canada. When you do your logging, is it necessary for the drill to stand by to clear the hole? You say you go into a lot of commercial holes. What is the situation there, do you have to follow right along

immediately behind the drill? Do you have to make special arrangements with the driller that he case the hole with plastic tubing, etc.?

Keys. More often than not, there is no drill rig at the site when we get there. We have had a project in Anchorage, Alaska, for the last two summers. These holes are in unconsolidated glacial fluvial sediments. Every well in the Anchorage area is cased to the bottom. Any uncased portion would cave in almost immediately. So the only techniques that we can use are radiation, gamma, gamma-gamma and various neutron logs. In the Anchorage area, we make as many contacts as possible as to where wells might be available for logging, city wells or private. We are constantly being called and we go out there and log them. For open holes, depending in which rocks it is drilled, we rent a hole in granitic rocks near Denver. This hole was drilled at least 15 years ago. Other holes, I have gone in the minute the drill stem was pulled out and the hole caved in before we got to the bottom. So it just depends on the rock material. Usually we do not have a driller there at the hole.

Collett. On developing these new techniques in logging, do geologists approach you on their requirement or is this a personal interest or is it part of the result of communication between the hydrogeologist and yourself?

Keys. I think that most of my contacts are personal. I might say that I act as a consultant on the staff of the Water Resources Division. Each of us covers a field in which we are expert. Now several thousand people in the Water Resources Division know about me and if they have a problem that might be solved by logging, they contact me. We would then go out and log the hole, help them in the interpretation of their logs. We may make suggestions as to which tools might be added to their logging equipment.

Collett. Well, I think I speak for all in this room, when I say that you should be highly commended for the work that you are doing in drill-hole logging techniques and interpretation.

Resistivity mapping by electromagnetic methods

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Abstract. A review is made of the electromagnetic methods for soil and rock resistivity mapping on a regional scale using ground and airborne techniques.

Field strength measurements from local broadcast stations in the 550–1600 KHz frequency range have been used to produce soil conductivity maps of Canada and the United States based on Norton's field intensity curves. U.S.S.R. authors have pointed out the feasibility of using radio waves for regional resistivity and geological mapping. VLF energy in the 10–30 KHz range from communication and navigation transmitters has recently been reported for mapping vertical conductors and faults.

The airborne AFMAG system which measures natural electromagnetic field energy in the audio frequency range has been employed for mapping large fault and shear structures. An example is given for the L. Abitibi-Noranda area to demonstrate the capabilities and limitations of this method.

In general, airborne electromagnetic systems have been applied to the search for ore deposits. These EM systems have been designed to minimize the effects of conductive overburden. However, a great deal of regional geological information can be obtained from these systems with or without modifications to the equipment. An example is given in which the conventional Rotary Field, Rio-Mullard and AFMAG systems were used to map changes in the overburden and rock contacts in Africa. Recently the Barringer INPUT system has been modified for mapping the resistivity changes in the overburden. An example is given in which INPUT was used to outline an aquifer in Manitoba. In France the INPUT system has also mapped the contact between fresh water and salt water.

Résumé. L'auteur présente une revue des méthodes électromagnétiques employées, tant au sol que dans les airs, pour la cartographie de la résistivité du sol et des roches à l'échelle régionale.

Des mesures d'intensité de champ, à partir d'émetteurs locaux dans la gamme de fréquences 550–1600 KHz, ont été utilisées pour obtenir des cartes de conductivité du Canada et des Etats-Unis fondées sur les courbes d'intensité de champ de Norton. Les auteurs russes ont souligné la possibilité d'utiliser les ondes radio pour cartographier la résistivité et la géologie régionales. L'emploi de l'énergie dans la gamme de 10–30 KHz, provenant d'émetteurs utilisés dans le domaine des communications et de la navigation, a été récemment annoncé pour la cartographie des conducteurs verticaux et des failles.

Le système aéroporte AFMAG, qui mesure l'énergie naturelle du champ électromagnétique dans la gamme BF. a été utilisé pour la cartographie des grandes failles et des structures cisaillées. La région du lac Abitibi-Noranda est donnée à titre d'exemple afin de démontrer les possibilités et les insuffisances de cette méthode.

En générale, les systèmes électromagnétiques aéroportes ont été utilisés dans la recherche de gîtes de minerais. Ces dispositifs électromagnétiques ont été conçus pour réduire au minimum les effets de la conductivité du mort-terrain. Toutefois, nombre de renseignements sur la géologie régionale peuvent être obtenus grâce à ces systèmes, avec ou sans modification de l'appareillage. A titre d'exemple, on utilise les systèmes classiques Rotary Field, Rio-Mullard et AFMAG pour dresser la carte des variations dans le mort-terrain et des contacts rocheux en Afrique. Récemment, le système INPUT de Barringer a été modifié pour permettre de cartographier les changements de la résistivité dans le mort-terrain. A titre d'exemple, on utilise l'INPUT pour délimiter une zone aquifère du Manitoba. On a également démontré en France que le système INPUT peut cartographier le point de contact entre l'eau douce et l'eau de mer.

For over three decades now, electromagnetic techniques have been studied and systems developed to detect subsurface conductors for locating sulphide ore. During these investigations, it has been realized that electromagnetic parameters of the host rock and the presence of layers of soil overlying the locality modify the response of conductive ores (Lowrie and West, 1965). The local resistivity of earth materials is a statistical concept but there is a relationship between this resistivity and various types of soils and rocks. The salt and water contents of soils and rocks have a strong influence on their resistivity (Kashprovskiy, 1963a).

Since 1948, geologists have been using magnetic data for mapping many geological structures. With advances in electromagnetic instrumentation and a better understanding of the influence of electrical properties of the ground on electromagnetic fields, a new era of electrical resistivity mapping is now being entered. Electromagnetic methods can detect conductive faults and shear zones, differentiate between various soil and rock types, and map sand and gravel deposits (possible aquifers) and fresh water–salt water contacts. Electrical resistivity maps can aid the mapping geologist by supplementing information acquired by other means. For the production of regional maps based on electrical parameters, methods are now being studied for acquiring data as efficiently and cheaply as possible and in a form that can be processed by computers and plotters. The purpose of this paper is to review those audio and radio frequency electromagnetic methods which appear suitable for electrical resistivity mapping on a regional scale.

Radio waves

Radio broadcast frequencies. The possibility of using radio frequency waves in geological investigations has interested geophysicists and other investigators of various countries since the 1930s.

In passing along the surface of the earth, some of the energy is refracted into the earth causing currents to flow. The intensity and distribution of these currents depends on the resistivity, dielectric constant and magnetic permeability of the upper layers of the crust. The structure of the radio wave at and above the surface of the earth is modified by the presence of the earth because of the loss of energy within earth and the fields associated with currents flowing in the earth.

The frequencies used in radio-wave transmission are generally much higher than those used in inductive prospecting methods. In Canada and the United States, broadcast stations cover the frequency range from 540 to 1640 KHz. The skin depth (depth at which the signal strength is reduced by the ratio 1/e) ranges from a few tens of feet in normally conductive rock to a few hundreds of feet in resistive rock at broadcasting frequencies (Keller and Frischknecht, 1966). For this reason, radio-wave measurements

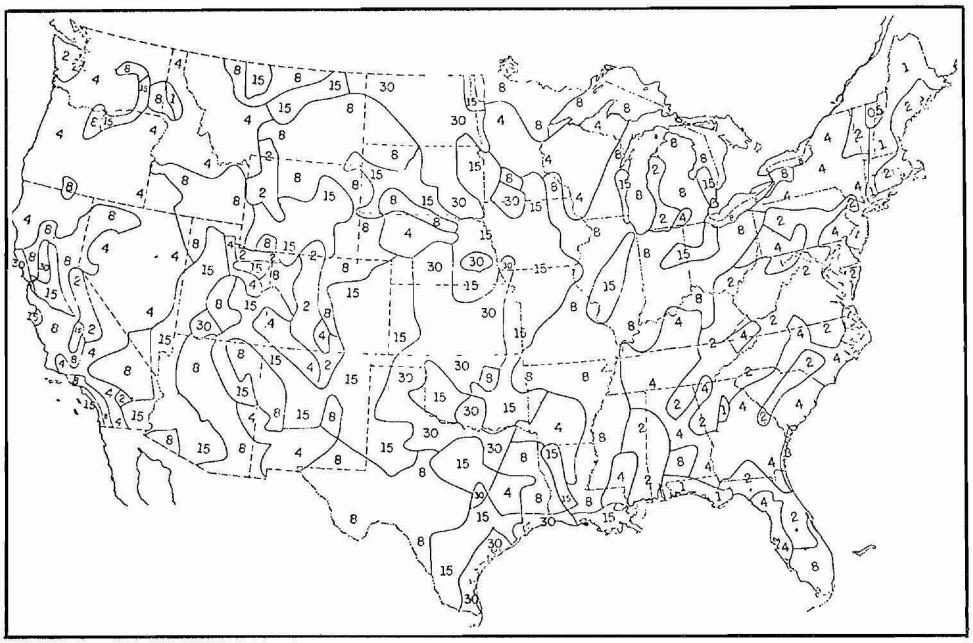


Figure 1. Estimated effective ground conductivity map of the United States using radio-wave field intensity methods (after Fine, 1954). Numbers on map represent estimated effective ground conductivity in millimhos per meter. Sea water is assumed to be 5000 millimhos per meter.

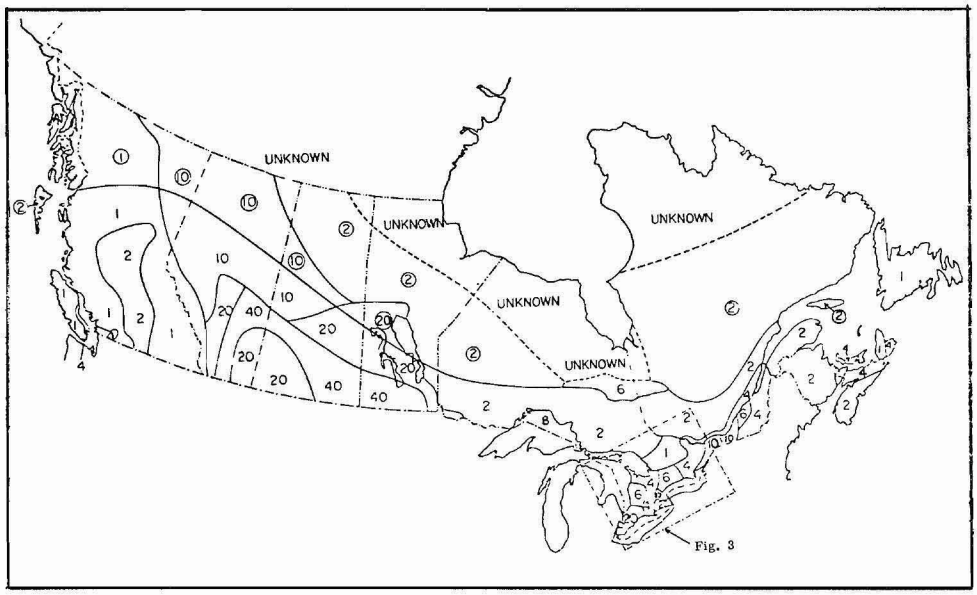


Figure 2. Provisional ground conductivity map of Canada using radio-wave field intensity method (after Ireland, 1961). Numbers on map represent average conductivity in millimhos per meter. Where conductivity values are circled, indicated value is based on limited data.

Table I. Conductivities for ground conductivity map of the United States.

(millimhos/m)

Average Conductivity	Range of Conductivities
0.5	.707
1	.707 – 1.414
2	1.414 – 2.828
4	2.828 – 5.657
8	5.657 – 10.95
15	10.95 – 21.21
30	21.21 – 45
5,000	Seawater

using standard broadcast stations as an energy source are best used for studying the electrical properties of the soil cover and overburden, and in areas where the soil is thin or resistive, the bedrock geology.

Norton (1941) has calculated a set of master curves for the decay of ground-wave field intensity with distance from the transmitting antenna. At a distance of up to 25 miles from the transmitter, the only signal observed at the earth's surface is the ground wave. Norton's curves were used by Fine (1954) to produce an 'effective' ground conductivity map of the United States, Figure 1. The values on the map are in conductivity units of millimhos per meter and each value represents a range of conductivities as shown in Table I.

Similarly, Ireland (1961) has produced a ground conductivity map using radial path field strength measurements from vertical

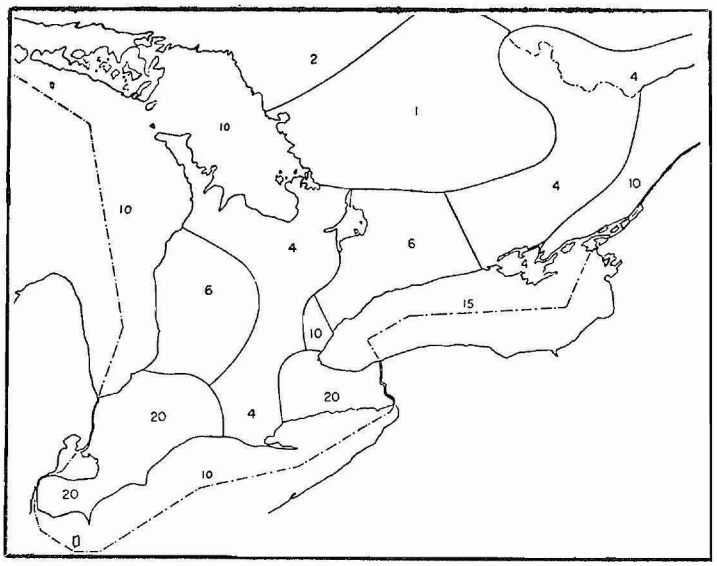


Figure 3. Provisional ground conductivity map of southern Ontario using radio-wave field intensity method (after Ireland, 1961). Numbers on map represent average conductivity in millimhos per meter.

antenna transmitter stations in the broadcast band frequency range. The ground conductivity map for Canada is shown in Figure 2 and that for southern Ontario, Figure 3. The units are in millimhos per meter and each value represents a range of conductivities as shown in Table II.

The conductivities for Canadian fresh-water lakes are given as follows (in millimhos/m): Lake Ontario, 15 (66.7 ohm-m); Lakes Erie and Huron, and Georgian Bay, 10 (100 ohm-m); and Lake Superior, 8 (125 ohm-m).

These conductivity maps have been adopted for use by the Federal Communications Commission in the United States and by the Canadian Association of Broadcast Consultants in Canada. Determining conductivity over land areas by using broadcast frequencies is subject to some approximations because of the effect of climate and terrain. Therefore, any conductivity map based on this type of information can give only a general conductivity picture. Admittedly, the Canadian map, in conjunction with the U.S. map, produces anomalous situations along the international boundary. This is because the two maps were drawn up by two different agencies evaluating information independently of one another. Other types of maps, such as soil maps, contain similar anomalies along the border (Ireland, 1961).

Table II. Conductivities for ground conductivity map of Canada.

(millimhos/m)

Average Conductivity	Range of Conductivities
1	– 1.414
2	1.414 – 2.828
4	2.828 – 4.889
6	4.889 – 7.746
10	7.746 – 14.14
20	14.14 – 28.28
40	28.28 – 56.56

In the U.S.S.R., practical methods of using radio-waves for geological mapping have been considered. Frolov (1961), and Khmelevskoy and Frolov (1959) have indicated the feasibility of radio-wave methods using the frequency range of 150–450 KHz for large-scale mapping and exploration. Their work verified the definite applicability of this method for mapping contacts of dissimilar rocks such as limestones, clay shales, sandstones and porphyrites and for detecting karst cavities in the northern Urals. They also were able to fix the contact between frozen rocks and thawed rocks. Kashprovskiy (1963a, 1963b) and Kashprovskiy and Kuzubov (1963), also of the U.S.S.R., have analysed measurements made by the radio-wave method and have shown that the derived data correlates well with soil groups. A conductivity map of the U.S.S.R. has been prepared by Kashprovskiy (1963a) but does not include the permafrost regions.

Fritsch (1963) of Austria has worked for many years on the influence rock formations on the propagation of radio frequency electromagnetic fields both on the surface and underground. In his excellent review paper, he shows a geoelectrical map of Austria based on the radio-wave method.

Very low frequencies (VLF). Although the earliest long-distance communications were performed with the aid of VLF (3-30 KHz) radio waves (Rugby 16 KHz during World War I), these frequencies fell into disuse in the middle 1920s. This was because Marconi and others showed that communication over long distances could be obtained more economically by the use of high-frequency (3-30 MHz) signals, although with lower reliability. This situation remained until about the 1950s when some of the attractive features of VLF communication, such as relative immunity to ionospheric disturbances, remarkably high phase stability, and relatively easy penetration into sea water began to outweigh the disadvantages for certain purposes (Crombie, 1964).

Although a great number of papers since 1950 have been written on VLF propagation characteristics, mostly dealing with the influence of the ionosphere, almost no attention has been given to the study of the effect of geological conductors on VLF waves. The Geological Survey of Sweden carried out some field studies in 1963-64 to investigate the possibility of utilizing VLF radio waves for ore prospecting (Paal, 1965). The results obtained showed clearly that VLF radio-wave measurements can be successfully used for prospecting with a ground instrument. Airborne applications were mentioned, although no airborne measurements were made. In 1966, Vaino Ronka, Geonics Ltd., put on the market a portable VLF ground instrument, EM-16, that is tunable to two VLF frequencies.

In 1966, the Geological Survey of Canada made some trial measurements with the EM-16 to determine if VLF waves could be used to detect geological structure. Becker (1967) chose an area over the Gloucester fault in the immediate vicinity of Leitrim, Ontario (Figure 4). The Paleozoic sediments on the downthrow side of the fault are considerably more conductive than the thin layer of dolomite and sandstone which overlies Precambrian granite and gneiss on the upthrow side of the fault (Wilson, 1946). The signal from station NAA, Cutler, Maine, is a 1000 kw transmitter operating at 17.8 KHz. This station, which is operated for communication with submarines, has a vertical antenna. The antenna current is thus vertical, creating a concentric horizontal magnetic field which propagates radially outward in all directions. When the magnetic field meets

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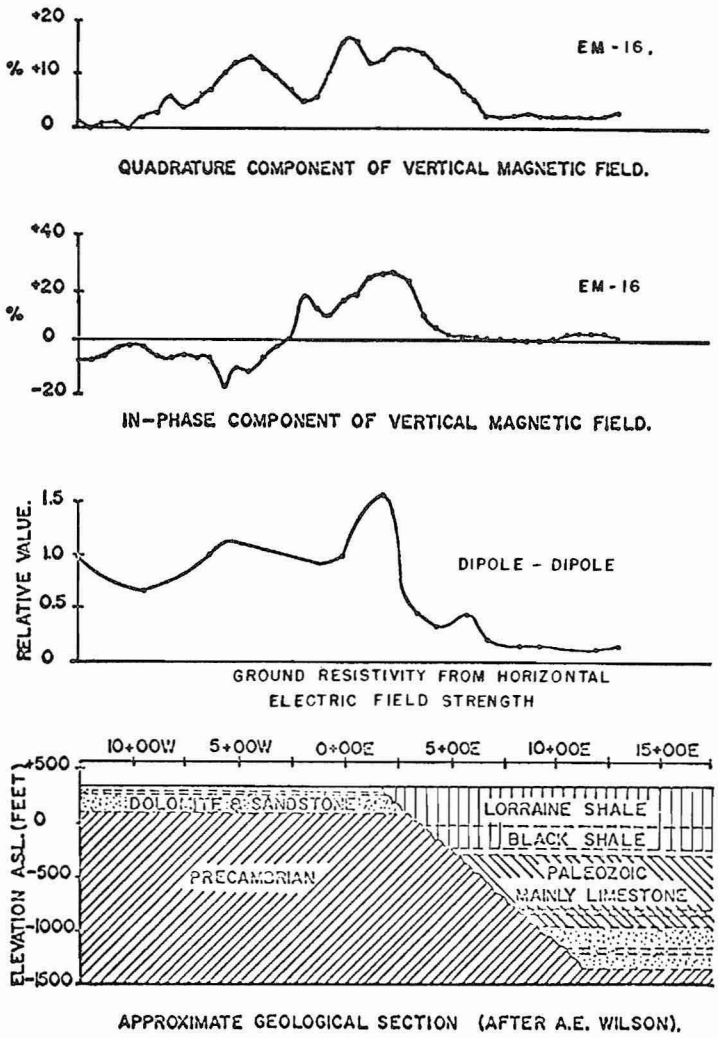


Figure 4. VLF field across Gloucester fault, Leitrim, Ontario.

conductors in the ground, secondary magnetic fields are reradiated from these conductors.

The EM-16 is simply a receiver which has two built-in ferrite-core receiving coils. One coil has a vertical axis and the other is horizontal. The signal from the vertical axis coil is first minimized by tilting the coil. The tilt angle is measured by a mechanical inclinometer in percentages. The remaining quadrature signal in this coil is balanced out by a measured percentage of the signal from the horizontal axis coil which has been shifted by 90°. The axis of the horizontal coil is kept parallel to the primary magnetic field of the VLF radio wave, which is at right angles to the direction of propagation from the station. Thus if the secondary magnetic signals from the conductor are small compared to the primary horizontal magnetic field, the tilt-angle is an accurate measure of the vertical real-component, and the compensation signal from the horizontal coil is a measure of the vertical quadrature component.

The two top curves in Figure 4 show the in-phase and quadrature component of the VLF signal over the Gloucester fault. The third curve made with a method using a dipole-dipole configuration, is analogous to the classical telluric current method. The equipment consisted of two modified Navy VLF receivers, which were adapted for measuring the horizontal component of the electric field vector. In the VLF range, the strength of this component varies proportionally with the square root of the resistivity. Both techniques indicate the presence of the fault zone. In addition, the electric field measurements show that shales on the downthrow side of the fault are about seven times as conductive as the dolomites and sandstones on the upthrow side.

AFMAG

The AFMAG method was first made generally available for geophysical prospecting in 1958 (Ward, *et al.*, 1958; Ward, 1959, 1960). AFMAG (an acronym for audio frequency magnetics), is an inductive electromagnetic method in which the source of energy is the natural audio frequency magnetic field arising from atmospheric discharges. It is classified as a dip-angle electromagnetic system with the transmitter removed to 'infinity'.

In the airborne version of AFMAG, a two-coil receiver is towed in a bird by an aircraft. The tilt angle of the natural magnetic field is measured electronically. Normally, the magnetic field is horizontal. However, when a conductive zone such as a fault, dyke or sulphide orebody is present, eddy currents are set up in the conductive zone which in turn reradiates a secondary magnetic field. This reradiated field has a vertical component which causes the receiver to record a non-zero tilt-angle locally on either side of the conductive zone.

In 1958 the American Metal Climax, Inc. flew the first extensive airborne AFMAG survey covering some 1500 square miles in the Lake Abitibi-Noranda area of NW Quebec and adjacent parts of Ontario. During the examination of this airborne data, Shaw (1962) came to the conclusion that there was a type of response that was distinctly very different from that considered to be of potentially economic interest, and which could be traced for long distances from line to line through the survey area. Detailed geology is available for large portions of the survey area, and these large AFMAG anomalies showed marked coincidence with several of the known major structural breaks.

A plot of the AFMAG response and the faults are shown for the Lake Abitibi-Noranda area in Figure 5. Flight lines are north-south and are spaced 1/4 mile apart. The black dots mark the cross-overs of the large anomalies.

The Cadillac, Horne, Here Creek and Davidson faults lie in areas of strong powerline interference, therefore the indications over these zones cannot be regarded as reliable. There is a considerable amount of powerline interference over the central part of the Porcupine-Destor fault and, although the zone does show up, the pattern formed is extremely complex.

The location of the conductive zone along the Lois Lake fault does not lie exactly over the mapped position of the fault, but about 600 feet to the south, along the andesite-agglomerate contact. The Lois Lake shear zone is known to carry abundant graphite and sulphides, and these may be concentrated along this contact zone rather than along the zone of maximum ground movement.

The Porcupine-Destor fault zone is known to extend for some 140 miles from the Timmins-Porcupine area to the edge of La Pause township at the eastern edge of the map area. This zone has associated gold deposits at various places along its length, has

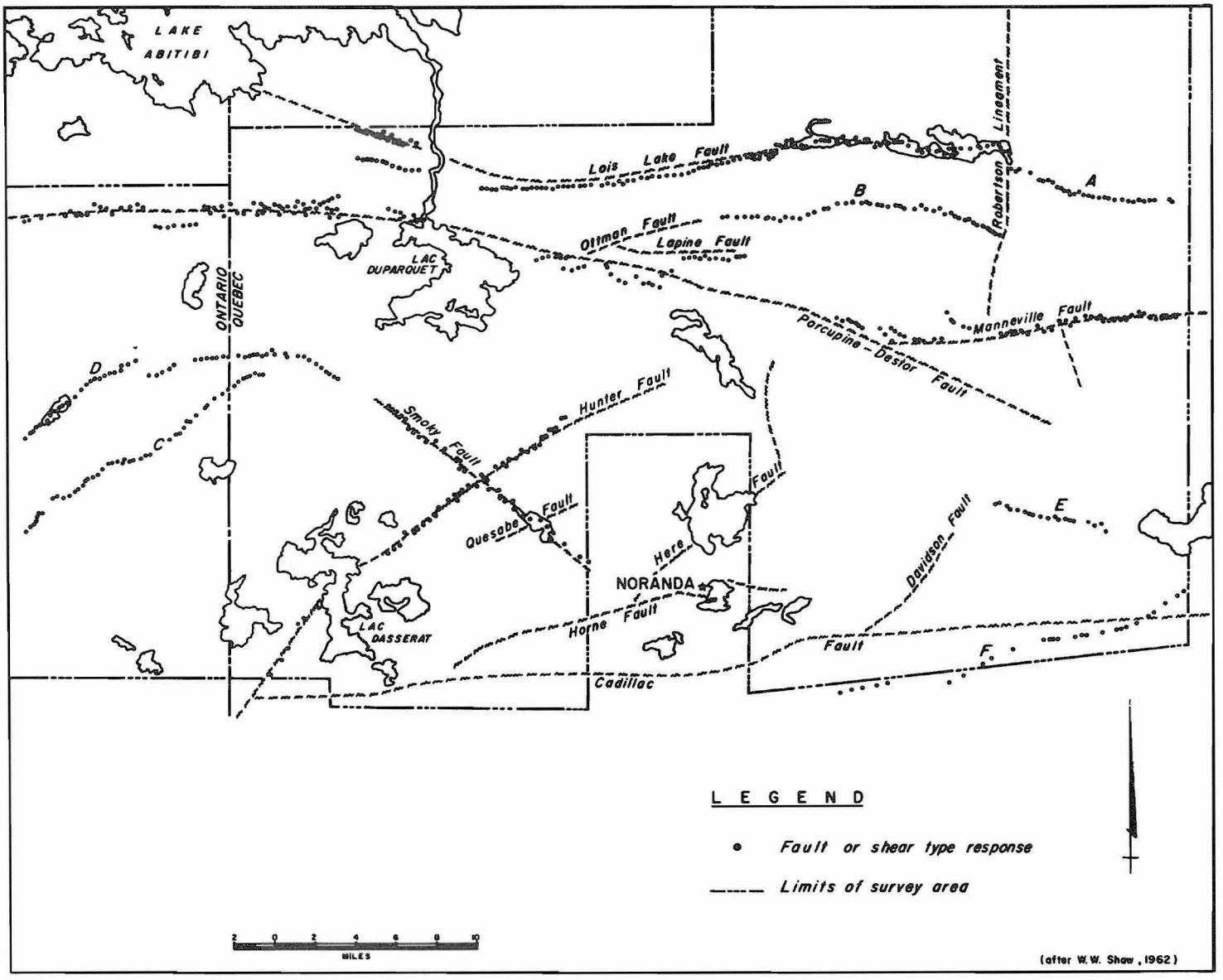


Figure 5. Fault and shear type AFMAG responses in the Lake Abitibi-Noranda area.

been very intensively drilled in search of new deposits, and is, therefore known in great detail. The AFMAG anomaly pattern over this zone is quite different from that over other features in the area, being extremely broken up in places, sometimes showing pronounced *en echelon*, offset and complex patterns.

Zone 'A' probably marks an extension of the Lois Lake fault. Zone 'B' reflects a possible continuation of the Ottman fault. Zones 'C' and 'D' trend NE-SW which is a common direction for the major discontinuities in this area.

Shaw (1962) suggests that in glaciated areas where highly variable overburden gives rise to interfering anomalies, it should be possible to obtain structural information by flying between 500 and 1000 feet above the ground. Ward, *et al.* (1966) point out that the repeatability of results over good conductors is generally good while over poor conductors it is notoriously poor. When the major axis of the inducing magnetic field is parallel to the fault, small or negligible dip angles result. On the other hand, the dip-angle response is often substantial when the major axis of the inducing magnetic field is normal to the fault. Thus in an area where a well developed conjugate fault system is suspected, mapping may need to be repeated at different times of the day or in different seasons of the year to trace all elements of the fault system. However, it has been Shaw's experience that deep-seated good conductors will always be indicated during the period from June to September in Canada, no matter the time of day.

AFMAG, Rio-Mullard and Rotary Field EM systems

In Africa, a comparison of three electromagnetic systems was made by Makowiecki, *et al.* (1964). Although the EM systems were compared for mineral detection, the results are important for mapping changes in resistivity in general.

The areas chosen for the comparison surveys are typical of a tropical region and the geology is fairly well known. The two

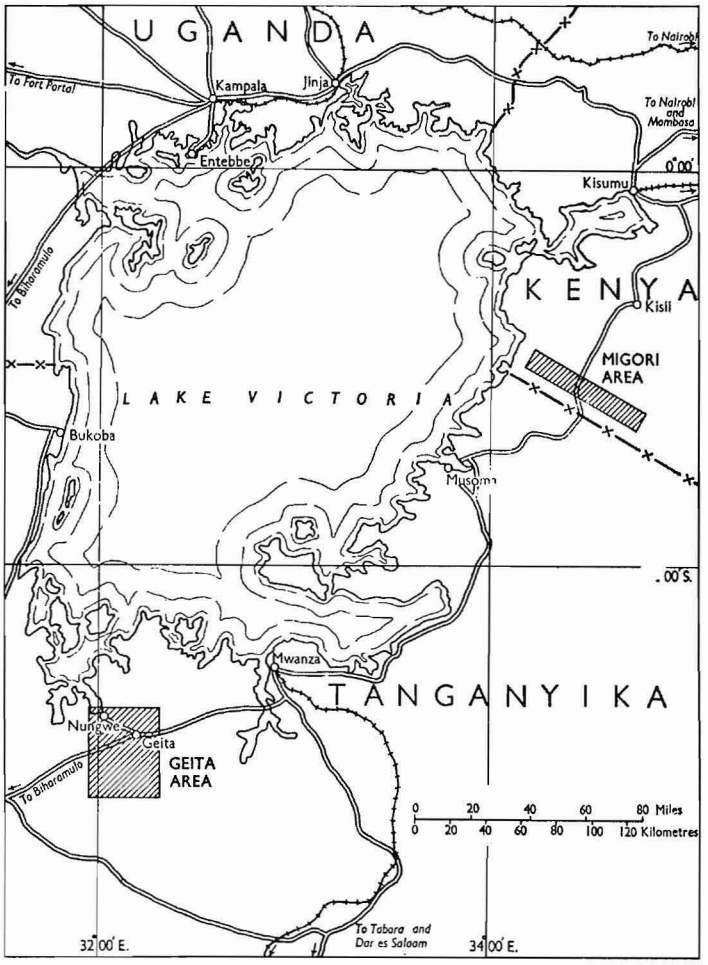


Figure 6. The Geita and Migori test areas (after Makowiecki, *et al.*, 1964).

areas are shown in Figure 6 and lie close to Lake Victoria. In the Geita area, which we shall discuss for this illustration, huge 'rafts' of metasedimentary and metavolcanic rocks of Nyanzian (Precambrian) age occur in a northern and southern belt with a large area underlain by granite between them. The Nyanzian belts, which are similar to the mineral-bearing 'greenstone' formations in Canada, consist of a series of metavolcanic rocks and schists and a banded ironstone suite which contains intercalated layers of tuff and porphyry and which has a regional E-W strike. All the rocks are folded and steeply dipping. The surface conductors are probably the decomposed zone of the Nyanzian rocks, alluvium and lakes.

In the Geita area, the flight direction was N-S or perpendicular to the regional strike. Over most of the area, flight-line spacing was 1/4 mile but it increased to 2 miles over the large area of granite in the central belt of the Geita area. The Rio-Mullard and AFMAG contractors employed strip film for flight path recovery, while the flight path for the Rotary Field survey was plotted during flight by the navigator who also marked the control fiducials on the record. There is a lack of identifiable features in the area so that the accuracy of the location of anomalies is poor for all three surveys.

Electromagnetic responses over lake, marsh and alluvium are illustrated in Figure 7. The Rio-Mullard traces are hardly affected by the conductive surface features. AFMAG responses tend to be confined to the edges of flat-lying conductors and especially to steeply dipping conductors. The Rotary Field responses are markedly affected by the surface conductors and show a general 'base shift' of both the real (R) and quadrature (Q) components which are correlateable to the surface features.

Similar conclusions are drawn from Figures 8, 9 and 10 which show the responses of the three EM systems over the southern Nyanzian belt and granites. In Figure 8, the Rio-Mullard system shows very little response over the Nyanzian belt. In Figure 9, the AFMAG system responds mainly to the contact zones between

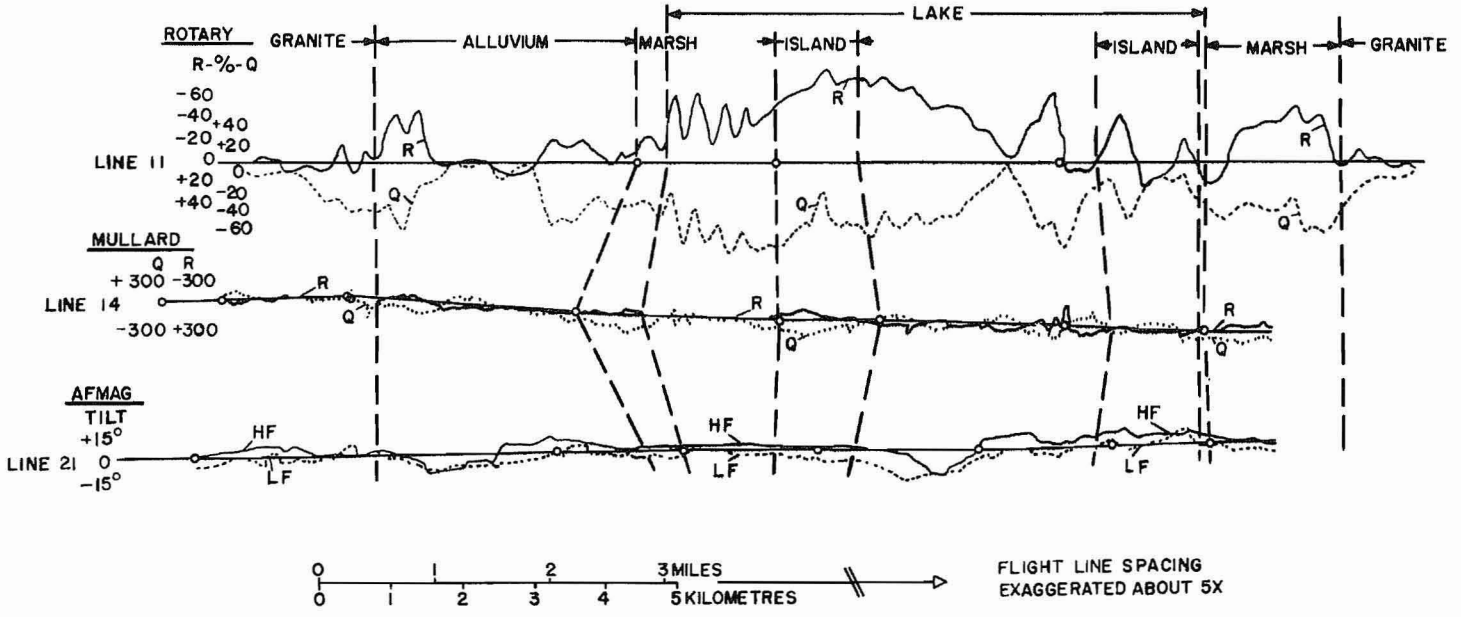


Figure 7. Aerial responses over lake, marsh and alluvium at Geita (after Makowiecki, *et al.*, 1964).

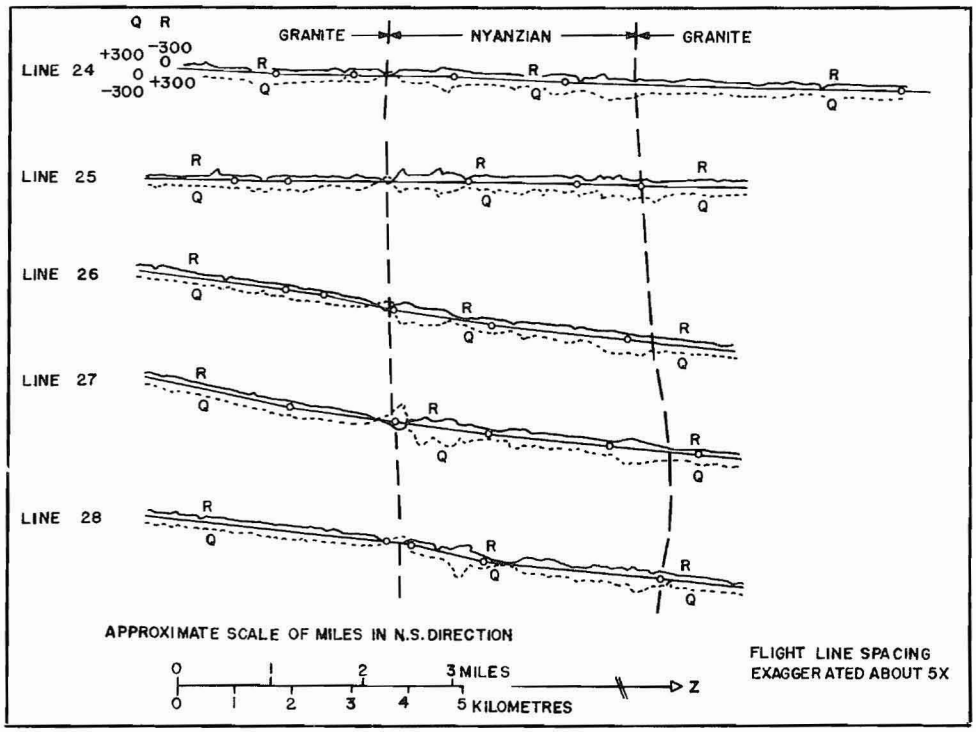


Figure 8. Rio-Mullard responses over southern Nyanzian belt, Geita (after Makowiecki, *et al.*, 1964).

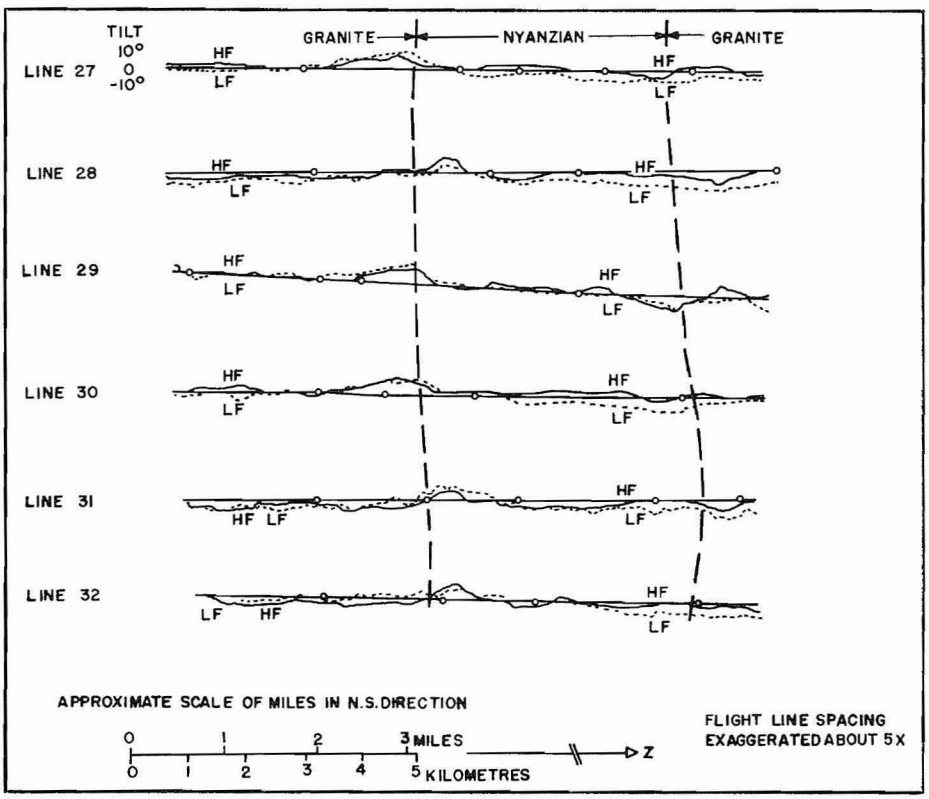


Figure 9. AFMAG responses over southern Nyanzian belt, Geita (after Makowiecki, *et al.*, 1964).

granites and the volcanic Nyanzian belt. In Figure 10, the Rotary Field traces are very active over the more conductive Nyanzian belt and are very inactive over the granites.

It must be pointed out that these tests were carried out with the purpose of assessing these EM systems for detection of sulphide ore. For instance, the Rio-Mullard system with its coplanar horizontal axis coil systems is designed to discriminate against noneconomic, flat-lying conductors and appeared to do this better than the other two systems. The Rotary Field system shows a high degree of response to highly conductive surficial

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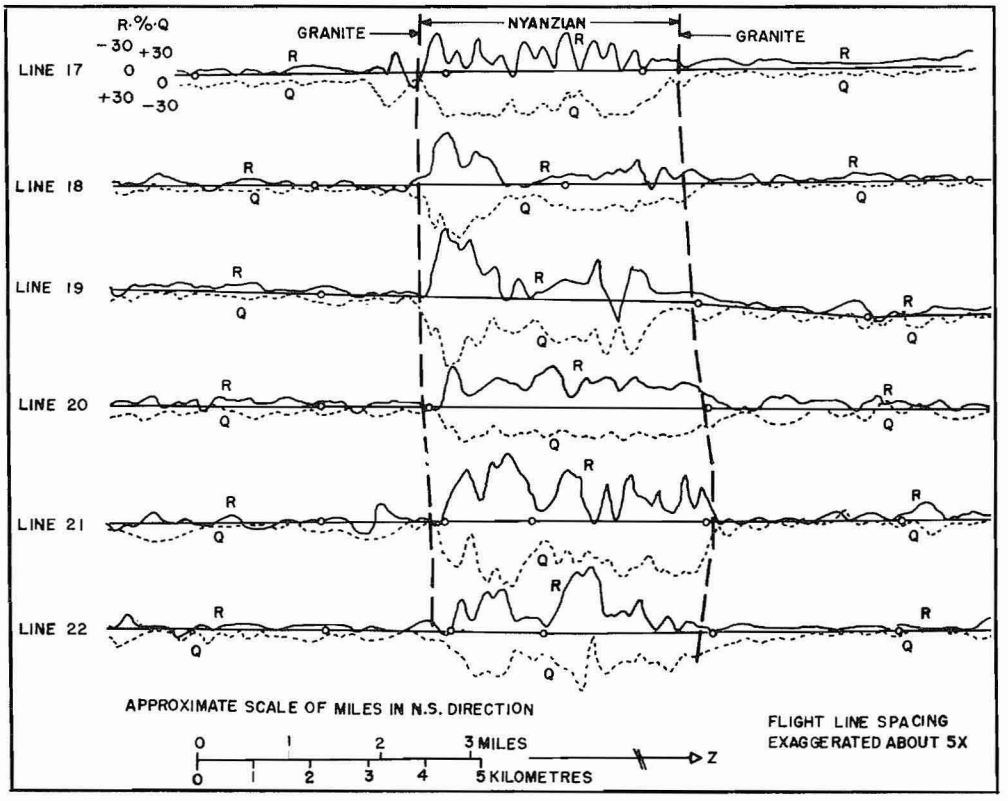


Figure 10. Rotary Field responses over southern Nyanzian belt, Geita (after Makowiecki, *et al.*, 1964).

Figure 11. DC resistivity map over Winkler aquifer, Manitoba.

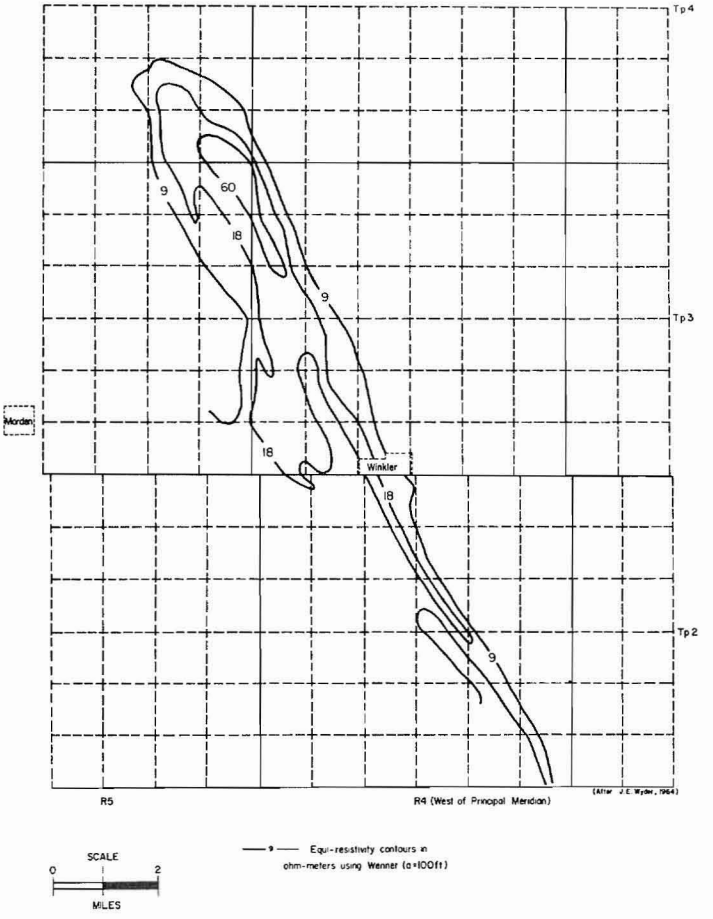
deposits and weathered material. The AFMAG system lacks resolution for small conductors but having a superior depth penetration capability, it is particularly effective in delineating long, poorly conducting fault and shear zones.

INPUT Mark IV

The INPUT (Induced Pulse Transient) system differs from continuous-wave airborne electromagnetic systems in that the current wave form consists of half-sine pulses of alternating polarity. The repetition rate is 288 pulses per second; the pulses have a base width of about 1.5 milliseconds and are spaced approximately 2.0 milliseconds apart. These current pulses are fed through a 3-turn transmitter loop mounted on a Super Catalina (PBY) aircraft.

The receiver, which is located in a bird that is towed on 500 feet of cable at a position approximately 250 feet behind and below the aircraft, records the decay of the transient electromagnetic fields generated in the ground by the pulses. The amplitude and shape of the transient is proportional to the conductivity of the ground. A good conductor produces a larger transient than a poor conductor. This transient voltage is sampled at six different periods following the termination of the pulse. The voltage output for each period is amplified and filtered by individual channels and recorded continuously on a multichart recorder. The height of the aircraft above ground is also recorded on the same record. The aircraft position is recorded by a 35 mm camera and each exposure is identified on the record by a fiducial mark.

In 1965, the Geological Survey of Canada had the INPUT Mark IV system flown over parts of townships 2, 3 and 4, ranges 4 and 5 (west of the Principal Meridian) in the Winkler area,



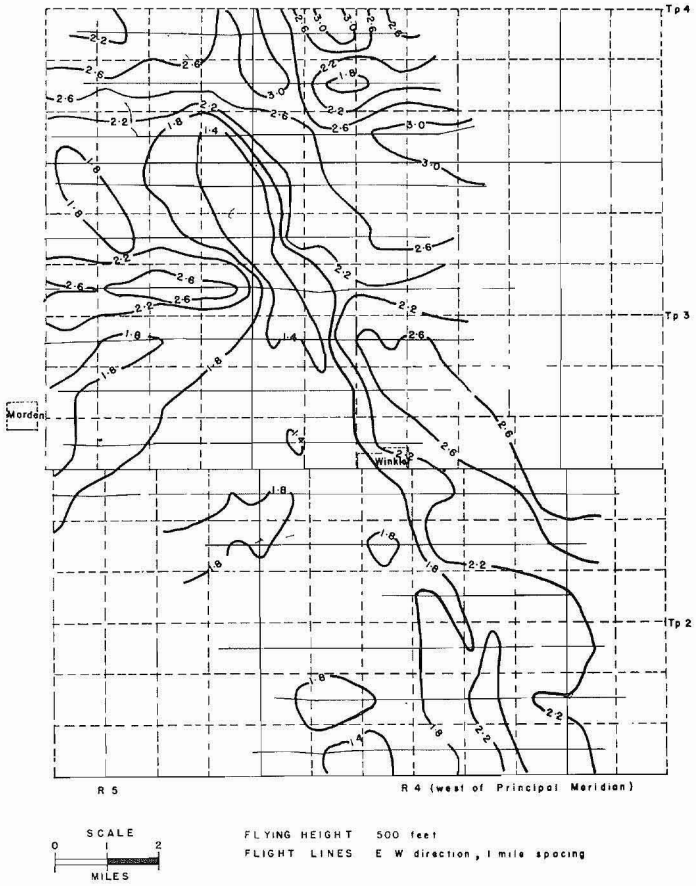


Figure 12. INPUT responses in volt-μsec over Winkler aquifer, Manitoba.

Manitoba to test whether a known aquifer could be outlined. For this test, the axis of the receiver coil in the bird was mounted vertically. The widths and positions of each of the six channels were set as shown in Table III.

Prior to the INPUT survey, a ground DC resistivity survey had been carried out over the Winkler aquifer by Wyder (1964) using a Wenner array with electrode spacing, a = 100 feet. The results shown in Figure 11 outlined what apparently is an isolated NW-SE trending sand and gravel aquifer about 15 miles long.

The INPUT response over the Winkler aquifer is shown in Figure 12. The contour lines in this figure represent the calculated total area under the transient curve in volt-microseconds. The flying height was 500 feet and the flight lines were E-W and were spaced 1 mile apart. The low response over the aquifer coincides very well with the ground DC resistivity high over the aquifer. The response is highest on the eastern side of the aquifer where the clay and till units are relatively thicker. On the western side of the aquifer, the response is intermediate in amplitude because of a thinning of the underlying clay and till. The bedrock shales (Ashville formation) which are quite conductive also are nearer to the surface. From the ground DC resistivity survey, the range of resistivity values of the different geological units are as follows (in ohm-meters): clay, 5–20; till, 10–30; sand and gravel, 40–600; and bedrock shale, 12–14.

Table III. INPUT channel settings for Winkler, Manitoba Survey.

Channel	No. 1	No. 2	No. 3	No. 4	No. 5	No. 6
Width in μ sec	300	300	300	300	300	300
Center: time in μ sec from end of pulse	250	450	650	1050	1450	1850

Although the airborne method does not give the apparent resistivity values as measured by the ground DC resistivity survey, the amplitudes of the transient responses do delineate areas of different resistivity contrasts. Changes in the lithology are resolved better by the INPUT survey. It should be possible using this method to delineate areas for future follow-up programs in Pleistocene geologic mapping and in drilling for water supplies and sand and gravel deposits in the Prairies.

Conclusion

A brief review has been made of electromagnetic methods which can be used to map changes in resistivity of geological rocks and soils. Radio broadcasting frequencies can be used to prepare maps that correlate fairly well with soil types. VLF methods, which have a deeper penetration, can be used to locate faults and shear zones in the bedrock and the method appears to be feasible for airborne reconnaissance mapping. AFMAG is being used by a major exploration company to map broad structural features such as faults and shear zones. This airborne technique appears to be very useful for broad reconnaissance mapping of geological features. Conventional airborne electromagnetic systems do hold some promise for geological mapping purposes, especially with some modifications. The INPUT system is successful in mapping changes in electrical resistivity of the Pleistocene deposits and for detecting sand and gravel (aquifer) deposits in western Canada.

Acknowledgment

I wish to thank Mr. W.W. Shaw, American Metals Climax Inc., for permission to use the AFMAG data in the Lake Abitibi-Noranda area.

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Discussion on Resistivity Mapping by Electromagnetic Methods, by L.S. Collett

Collett, L.S. The Geological Survey of Canada's philosophy behind reconnaissance airborne electromagnetic mapping technique is to assess all suitable systems in regard to their usefulness in interpreting geological structure, bearing in mind the cost of carrying out these surveys.

Hall, D.H., University of Manitoba, Canada. I would like to commend this work very much. Once you can make a map using an airborne technique, it really speeds up the job of the geologist.

Collett. Our next step with INPUT is to study the transient data theoretically and to record the analogue data digitally in an industry-compatible format. This will allow the data to be handled and processed by the computer. We are going to 12 channels in our next tests to prove whether the bumps that we sometimes see in the transient decay are real or whether it is an instrumentation problem. It usually occurs in one channel, either channel 4 or 5.

Utard, M., Geoterrex CGG, France. I think that this is due to instability and distortion from one channel to the other such as phase shift and temperature drift.

Giret, R., Compagnie Générale de Géophysique, France. In the 6-channel system, there is some overlap in the channels, won't there be a lot of overlap in your 12-channel system?

Collett. In the 6-channel we did not have any overlap since we did not use the standard gate settings. Also in the 12-channel system there will be no overlap. The axis of the coil in the receiver is vertical which is also not the standard orientation.

Utard. The transient amplitude in your survey is not very large, perhaps this was due to the resistivity and larger thickness.

Wyder, J.E., Dept. Energy, Mines and Resources, Canada. The ground resistivity is not as high as it might appear. We now know that the water in the lower sand layer in the one example is salty. It would appear that the resistivity in this layer is approximately the same as the overlying clay and till.

Collett. Between the time of this INPUT survey and the time of the survey that was conducted in the Rhone River delta, the transmitter power had been doubled.

Utard. By comparing some of your ground DC resistivity work with the INPUT anomaly in the NE corner of your map, you do not get much indication. You should have adjusted your electrode spread to fit the ground condition. The airborne method is 'self-adjusting' over the large sampling time because it covers a wide range of frequencies.

Giret. Too often ground resistivity work is done for water survey with the wrong distance between electrodes. It should be adapted to the depths. Some of the trouble that we encounter with clients and geologists comes from previous surveys carried out with the wrong electrode spacings.

Wyder. I would like to answer that by saying that the two profiles that you see there are two points off a 12-point vertical sounding. They are the best ones that we have. I did vertical profiling every half mile and in some cases every quarter mile. If I get a chance to go back into the area, I will change the program. I just did not have enough time this year to do a complete job.

Utard. Even a spread in one part of the area is not the right one for another part, I realize this.

Baudoin, P., BRGM, France. From the ground survey, can you assess the change in resistivity in the first layer?

Wyder. Do you mean the lateral change in the DC resistivity measurements? Outside of the aquifer, we went 6 miles east and 4 miles west working on 8.3 ohm-meters. The variation was less than 5%. Readings were taken at every 300 feet.

Hurr, T., U.S. Geological Survey. Did you do this before or after your airborne survey?

Wyder. This example here was done after, and we only had two days in which to carry it out.

Gendzwill, D., Saskatchewan Research Council, Canada. There is an application of electromagnetic methods which you neglected to mention which is the use of ground electromagnetic equipment of the variety used by mining prospectors. An example is in the New Brunswick area, where large areas of the Bathurst mining district were covered with horizontal- and vertical-loop EM equipment. This area consisted of acid and basic volcanics, and graphitic sediments, the latter almost invariably picked up by horizontal- or vertical-loop EM methods. Generally, the horizontal loop method was more precise. Geologists

mapping this area would always refer to EM maps before deciding on where they would place the contact of the sediments. This last year we tried the same method in the glacial deposits in Saskatchewan. An area near Saskatoon which has been covered in detail by resistivity was selected as a test site. We carried out horizontal-loop EM at two different frequencies and three different coil spacings and discovered that there is a correlation between DC resistivity measurements and electromagnetic response. As yet, we have not interpreted this quantitatively but preliminary charts that I have drawn up in areas of low resistivities show that the EM definitely does respond. I think that this work in Saskatchewan is significant after looking at your airborne EM results. It is useless to try to find thin layers of deeply buried sand by surface geophysics methods. The extent of the use of resistivity method with grounded electrodes is to detect shallow sand deposits of the aquifer type. Why not use EM? It has good penetration and it may be very useful to map the shallow extent of sands and gravels.

Collett. This is actually the next step that we are planning to do in our ground follow-up work using a broad frequency range.

Becker, A., Geological Survey of Canada. I would like to add to Mr. Collett's comments. We tried this actually using conventional EM equipment over a fault where we know we get a good airborne anomaly. We get a marginally small ground EM anomaly around that of operator error. This is because of the difference in orders of magnitude of the sensitivity between the airborne and ground systems. One instrument that seems to be well suited to this type of measurement is the Boliden Slingram which has a 10 ppm sensitivity. It is the only ground EM that I know of which has that kind of sensitivity. Dr. Parasnis tells me that there are great operational difficulties in carrying this unit over rough terrain.

Meneley, W., Saskatchewan Research Council. What would it cost to do an EM survey over Saskatchewan with an area of 100 to 150,000 square miles?

Collett. At half-a-mile spacing, that is 300,000 line miles.

Giret. $6,000,000.

Meneley. This is a point when we are talking about reconnaissance methods for groundwater as applied to mineral exploration. An orebody in mining is a very small target when compared to even a small aquifer. The Winkler study is already, in my mind, a very detailed study, not a reconnaissance investigation at all when you are dealing with only a few townships. This is one of the misconceptions in dealing with geophysical methods in groundwater: that you should have information, maybe not precise, but covering very large areas.

Giret. I think for a reconnaissance survey it is not necessary to go out to a wide spacing.

Collett. It certainly depends on the problem.

Meneley. $3,000,000 would be quite adequate to pay for the compiling of the geological mapping carried out in Saskatchewan over the past 8 years. I think in this particular context I would sooner take the regional geological mapping rather than an airborne EM map of Saskatchewan.

Giret. I think that both are needed.

Baudoin. I just want to say that EM methods respond to conductivity. In fact, what you can see by EM is lateral changes in the electrical resistivity of the conductive overburden.

Meneley. Many of our aquifers in western Canada are not highly resistive because of the salinity of the waters. Mr. Wyder has found out in Manitoba that our aquifers are highly mineralized and hence may not be highly resistive.

Collett. I can say that any new methods that we look into are costly at first. We have to experiment until we find an economical way to carry out reconnaissance surveys.

Etude par prospection électromagnétique aérienne d'un contact eau douce - eau salée dans le delta du Rhône

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Bureau de Recherches Géologiques et Minières
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M. Utard

CGG Geoterrex
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Résumé. Les auteurs exposent les résultats d'un test de prospection électromagnétique aérienne (méthode INPUT) effectué par le Bureau de Recherches Géologiques et Minières sur un contact eau douce – eau salée partiellement étudié au sol par sondages électriques. Il s'agit d'un problème complexe à trois et parfois quatre terrains, dont le premier et le troisième sont conducteurs. Il y a une bonne corrélation entre les cartes d'amplitudes des anomalies électromagnétiques et les cartes de conductance longitudinale établies au sol; les principales zones à eau douce y sont mises en évidence et délimitées. Les auteurs observent aussi, et la théorie permettait de s'y attendre, que la comparaison des cartes des amplitudes mesurées à des temps plus ou moins longs après la fin de l'impulsion primaire permet en certains cas de distinguer les variations de résistivité ou d'épaisseur du substratum conducteur, dont le toit est de 50 à 100 m de profondeur, de celles qui ont lieu dans les formations superficielles.

En conclusion, l'électromagnétisme aéroporté fournit ici une couverture d'intérêt comparable à celui qu'auraient deux cartes de résistivité correspondant à deux profondeurs d'investigation différentes. L'interprétation reste encore purement qualitative; elle doit être contrôlée et précisée par des travaux au sol.

Abstract. The results of an airborne INPUT survey on a salt water - fresh water contact previously studied by ground methods are presented. The geology consists of three to four layers, the first and third layer being conductive. The correlation between contour maps of the amplitude of EM anomalies and ground resistivity maps is good and evidence is given of detecting the main fresh-water zones. An interesting fact is that it is possible, in some cases, from the comparison of the various EM amplitude contour maps corresponding to measurements at various times after the end of the energizing pulse, to approximately determine whether apparent conductivity changes are caused by changes in shallow layers or at depth.

More generally, EM maps established from a multitime or multifrequency EM system appear to be as interesting as some conventional resistivity maps obtained with various electrode spacings. The interpretation however is purely qualitative and has to be controlled by ground geophysics.

Les techniques de l'électromagnétisme aéroporté, initialement mises au point en vue de la détection de corps conducteurs, amas de sulfures, par exemple, dans un environnement non conducteur, sont devenues assez sensibles pour que les réponses électromagnétiques soient mesurables même sur les formations médiocrement à très médiocrement conductrices rencontrées dans de nombreux problèmes hydrogéologiques. La cartographie de ces réponses fournit une carte de ≪conductance apparente≫, où les poids relatifs des terrains superficiels et plus profonds dépendent des caractéristiques du système utilisé: distance émetteur – récepteur, sensibilité, fréquence, résistivités même des formations survolées. Certes, le problème théorique consistant à calculer une de ces conductances apparentes pour des distributions données des conductivités dans le sol n'est pas complètement résolu. Quant au calcul inverse, il n'est possible que dans le cadre des hypothèses les plus simples. Il n'en reste pas moins que l'on peut établir, par géophysique aéroportée, et sur des terrains assez conducteurs, des cartes électriques d'intérêt comparable à celui des classiques cartes de résistivité apparente. Ce fait est déjà nettement mis en évidence par les résultats du test effectué en 1965 par la Commission géologique du Canada (Collett, 1966), avec un système INPUT, dans le Manitoba, où des chenaux de graviers aquifères dans un niveau ailleurs argileux apparaissent comme des zones de réponse minimum sur les cartes électromagnétiques.

Nous exposons dans le présent article les résultats d'un autre test effectué en mars 1966 par le Bureau de Recherches Géologiques et Minières, également avec un système INPUT, dans le Sud de la France. Le problème consistait à étudier le contact entre des eaux douces et des eaux salées, caractérisées par un fort contraste de conductivité. Les cartes électromagnétiques établies sont présentées ici et confrontées à la géologie et aux résultats des travaux géophysiques au sol.

Notons que si l'on mesure les réponses électromagnétiques dans un domaine de fréquence ou de temps suffisant (et c'est le cas de l'INPUT), il est théoriquement possible d'en tirer des indications sur la distribution verticale des conductivités dans les terrains survolés. Les méthodes telles que l'INPUT ou les méthodes à plusieurs fréquences de travail fournissent ainsi une information comparable à celle qui serait contenue dans des cartes de résistivité établies au sol avec plusieurs longueurs de ligne. Dans l'état actuel des choses, on ne sait malheureusement pas exploiter toute cette information; une interprétation complète, tenant compte en particulier de l'altitude de vol, ne pourra être effectuée qu'au moyen de calculateurs, après que l'étude théorique du problème aura été complétée, et que les techniques électromagnétiques elles-mêmes auront été encore améliorées. D'ores et déjà pourtant, des considérations simples peuvent guider de façon utile l'interprétation des résultats fournis par les systèmes existants qui mesurent la réponse électromagnétique à

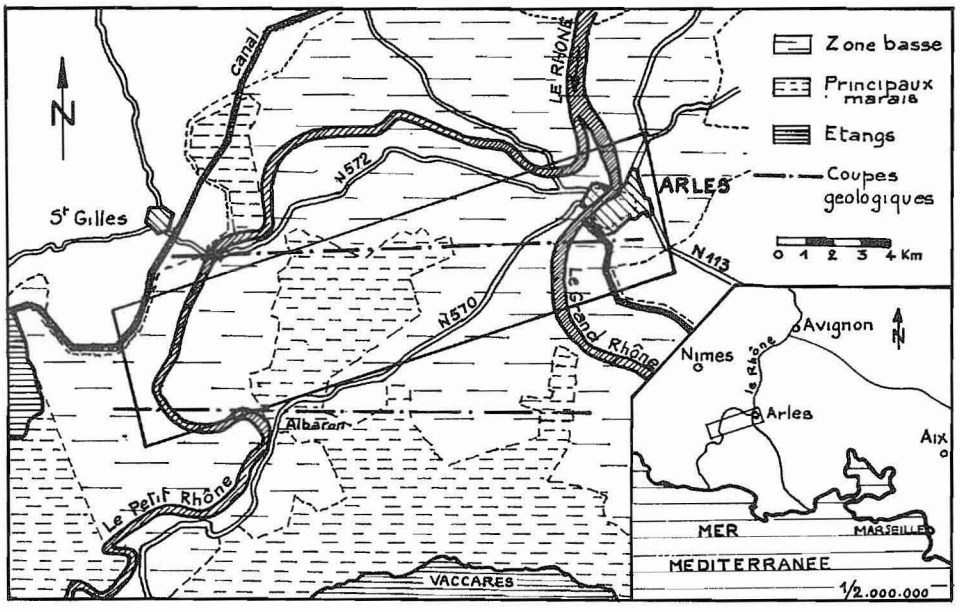


Figure 1. Plan de situation.

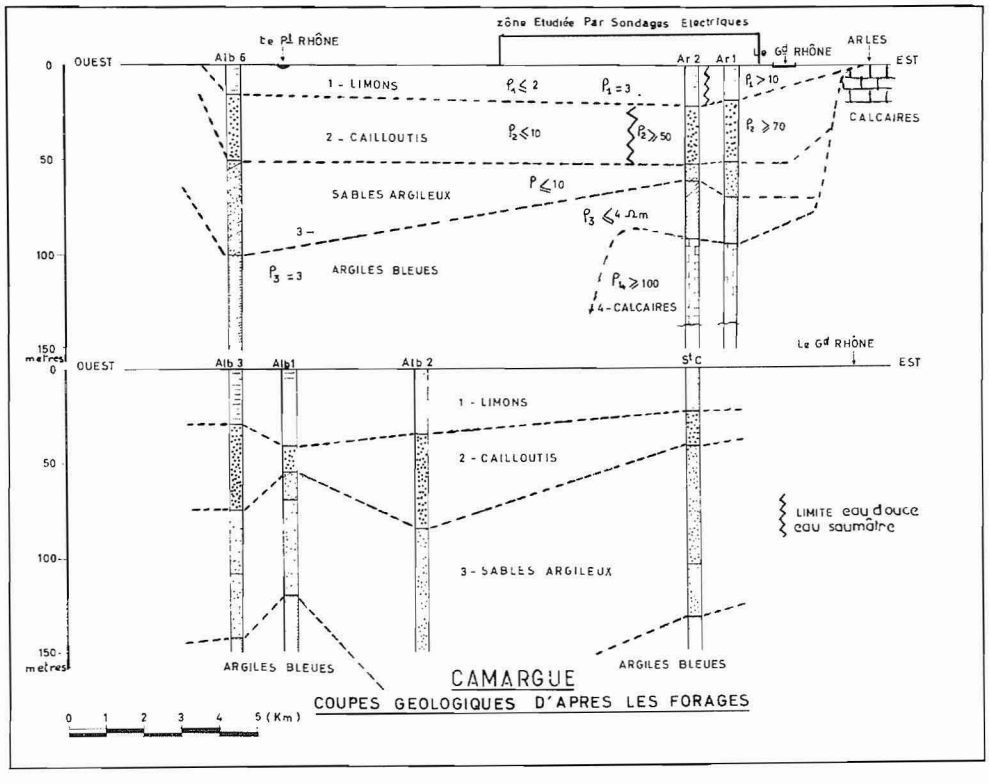


Figure 2. Coupes géolcgiques est-ouest.

plusieurs fréquences (technique fréquentielle) ou à plusieurs temps après la fin de l'impulsion primaire (technique transitoire). Dans un contexte donné, des changements de résistivité dans un terrain superficiel affectent surtout la réponse aux fréquences les plus hautes, alors que des changements de résistivité dans les terrains profonds affectent relativement plus la réponse à basse fréquence. De même, en régime transitoire, les variations de résistivité dans les terrains les plus profonds affectent le plus d'amplitude de la réponse longtemps après la fin de l'impulsion primaire. Encore la notion de profondeur reste-t-elle relative: la conductivité des terrains joue sur l'échelle des distances, de plus la profondeur d'investigation est limitée, et ce de façon variable suivant les systèmes utilisés.

Certaines des possibilités d'une telle analyse sont également illustrées par le cas présenté ici.

Exécution du test

Géologie et hydrogéologie. La zone couverte par le test est située en Camargue, dans la partie Nord du delta du Rhône, Sud de la France (figure 1). La plaine de Camargue est extrêmement plate

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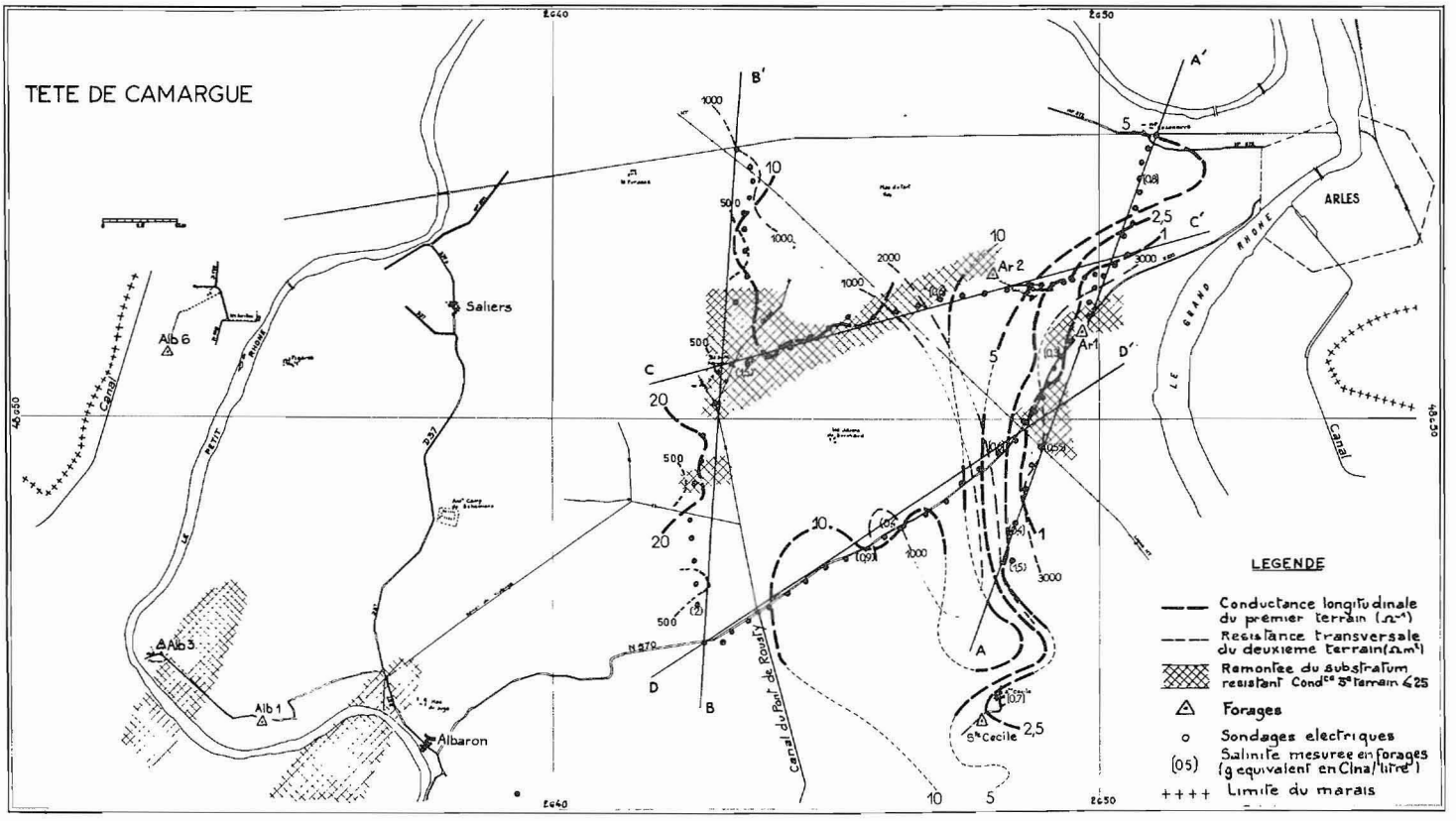


Figure 3. Synthèse des résultats des travaux géophysiques au sol.

et marécageuse, l'altitude entre les deux bras du Rhône étant comprise entre 1 et 6 mètres au-dessus du niveau de la mer. La nappe aquifère, très peu profonde, est en équilibre avec le Rhône. L'origine des eaux saumâtres et salées est assez mal connue; il est certain qu'à l'époque actuelle, les phénomènes d'évapotranspiration continuent à jouer un certain rôle dans la formation d'eaux saumâtres superficielles. L'alimentation en eaux douces est assurée essentiellement par le Rhône lui-même, la nappe des cailloutis pliocènes de la Crau et l'impluvium propre de la plaine (Durozoy, inédit).

La structure du bassin est partiellement connue grâce à des forages hydrologiques et pétroliers ainsi qu'à une série de sondages électriques effectués en 1965 par le B.R.G.M. (Stanudin, 1965) pour l'étude de l'alimentation en eau des agglomérations de la côte méditerranéenne.

La coupe géologique type est la suivante (figure 2):

Premier terrain: 20 à 40 mètres de limons gris bleuâtres

Second terrain: 30 à 60 mètres de cailloutis aquifères (Villafranchien)

Terrains	Résistivité (ohm-m)	Conductance horizontale (ohm $^{-1}$)	Résistance transversale (ohm-m^2)
1 (limons)	$0.7 < \rho_1 < 25$	$0.8 < C_1 < 20$	
2 (cailloutis)	$10 < \rho_2 < 100$	$0.3 < C_2 < 4$	$500 < R_2 < 3500$
3 (sables argileux et argiles)	$2 < \rho_3 < 15$	$10 < C_3 < 100$	

Troisième terrain: 0 à 500 mètres de sables argileux (Astien) et argiles bleues marines (Plaisancien)

Quatrième terrain: localement des calcaires jurassiques (faciès Urgonien) remontent à moins de 100 mètres sous la surface du sol (zones en hachures sur la figure 3). Ces calcaires sont affleurants à Arles. Dans l'ouest, le quatrième terrain, le plus souvent très profond, est constitué par des molasses miocènes.

Données géophysiques. Du point de vue géoélectrique, on est en présence d'un cas assez complexe à trois ou quatre terrains alternativement conducteurs et résistants. D'après les sondages électriques, on peut déterminer avec une assez bonne précision l'épaisseur h_1 et la résistivité ρ_1 du premier terrain (limons), le produit de l'épaisseur par la résistivité du second terrain (résistance transversale $R_2 = h_2\ \rho_2$), et faire une évaluation très approximative du produit de l'épaisseur par la conductivité du troisième terrain (conductance horizontale $C_3 = h_3\sigma_3$).

L'échelle des paramètres géoélectriques (résistivité, conductance horizontale, résistance transversale) définie d'après l'analyse des diagrammes de sondage électrique est donnée dans le tableau. L'examen du tableau montre que la conductance horizontale des différents terrains varie dans un large domaine, celle des premiers et second terrains variant surtout du fait de leur résistivité et celle du troisième surtout du fait de son épaisseur. De plus, la conductance horizontale du second terrain (aquifère) reste généralement faible devant celle des terrains situés au-dessus et au-dessous. On pouvait donc s'attendre à des réponses électromagnétiques assez contrastées, mais a priori difficile à analyser étant donné la complexité relative du schéma proposé.

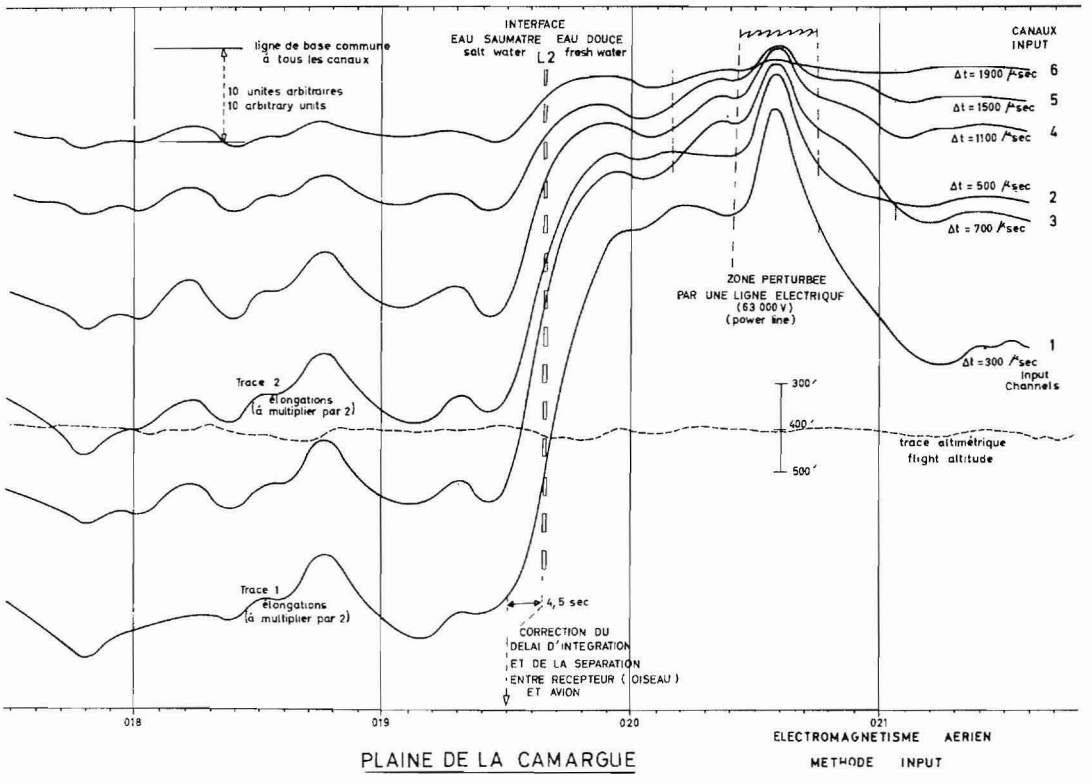


Figure 4. Facsimilé d'enregistrement INPUT.

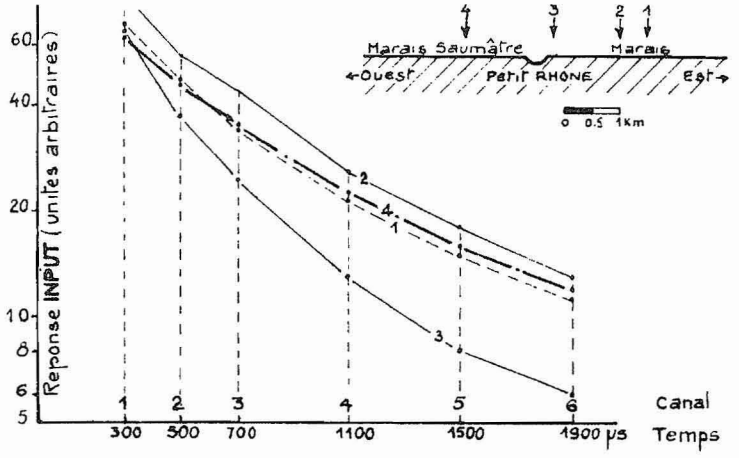


Figure 5. Courbes de décroissance, cas du Petit Rhône.

L'ensemble des résultats des travaux au sol est rassemblé sur la figure 3. Les conditions de terrain n'avaient pas permis, en pratique, de réaliser une couverture homogène par sondages électriques; aussi nous sommes nous laissé influencer, pour certains détails du contour, par les résultats électromagnétiques.

Campagne d'électromagnétisme aéroporté. Une seule sortie de 3 heures 50 effectuée le 2 mars 1966 a suffi pour le survol, à une altitude moyenne de 400 pieds, d'une bande de 20 kilomètres sur 4.5 kilomètres, au moyen de dix profils espacés de 500 mètres, orientés WSW-ENE, et d'une traverse nord-sud. L'avion, un Catalina Super-Canso affrété par Barringer Research Limited, était équipé du système INPUT MK V à six canaux, enregistrant l'amplitude de la réponse transitoire à des temps d'échantillonnages moyens de 300, 500, 700, 1100, 1500 et 1900 microsecondes après l'instant de coupure de l'impulsion primaire. La bobine détectrice installée dans l'oiseau était du type standard à axe horizontal, parallèle à la ligne de vol, dispositif qui ne donne pas un couplage optimum dans le cas de conducteurs horizontaux. Aucune spécification particulière n'avait d'autre part été imposée pour ce test (Bonniwell, 1966). On pouvait en effet considérer que, vue la topographie très plate de la Camargue, on ne serait pas gêné par de faibles variations de l'altitude de l'avion.

Nous ne rappelons pas ici le principe de la méthode qui a déjà fait l'objet de plusieurs publications (Barringer, 1962, 1963).

Résultats

Enregistrement INPUT. Un exemple d'enregistrement INPUT est présenté sur la figure 4. Les différentes traces analogiques correspondant aux six canaux INPUT ont été calées au début du vol sur une ligne de base commune. Un contrôle du zéro effectué en cours de vol montre une légère dérive dont il a été tenu compte. Les déflexions sont mesurées en unités arbitraires. (La sensibilité a été réduite de moitié sur les traces 1 et 2 pour lesquelles la déflexion est maximum. On lit donc de haut en bas sur les enregistrements les traces 6, 5, 4, 2, 3 et 1.) Le brusque gradient (ligne L 2) affectant toutes les traces marque le contact entre une zone à eaux douces (faibles déflexions) et une zone à eaux salées (fortes déflexions). Sur la partie droite de l'enregistrement, on observe une perturbation au passage d'une ligne à haute tension (63,000 volts). Cette perturbation a été lissée à la main; il peut en résulter une légère distorsion des cartes d'amplitudes de la réponse électromagnétique.

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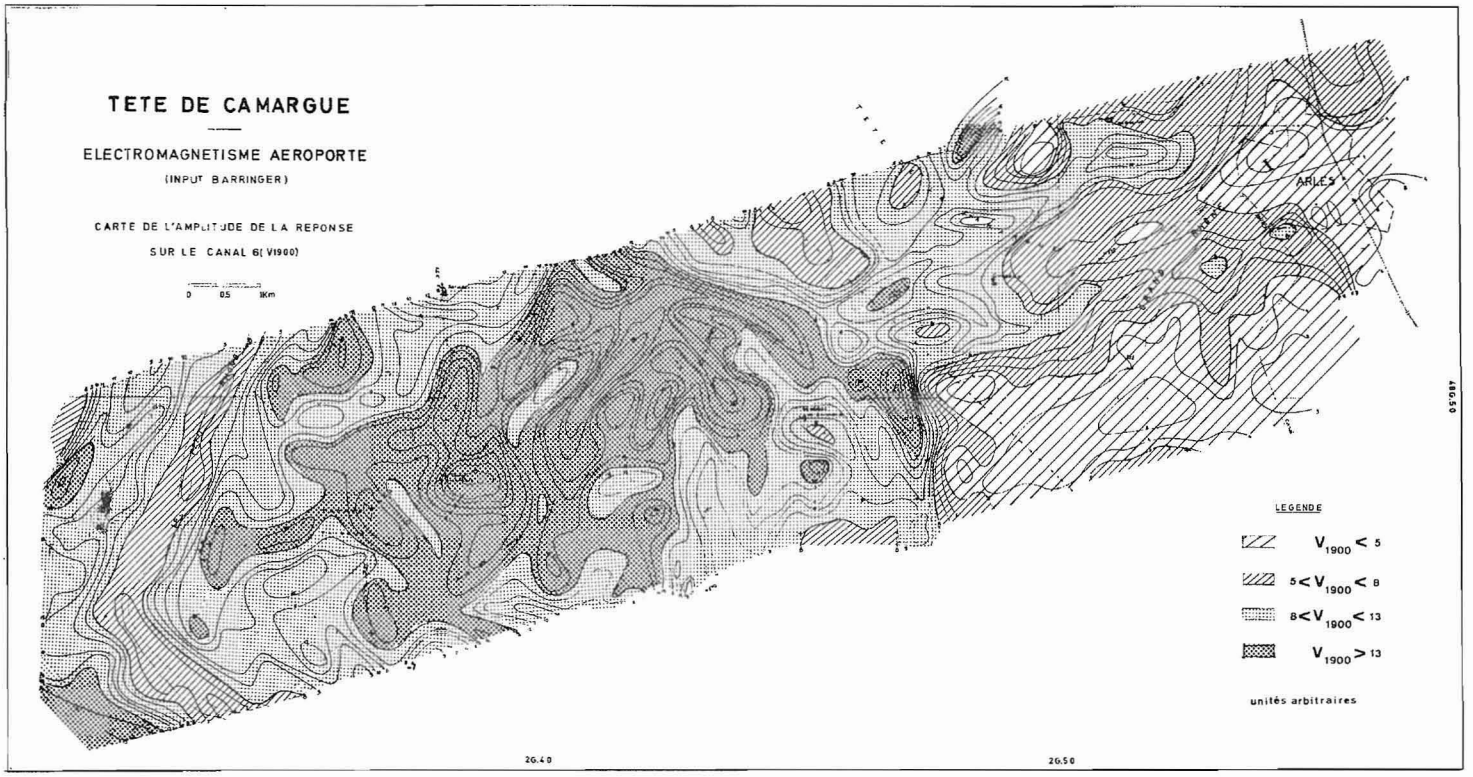


Figure 6. Amplitude de la réponse INPUT sur le canal 6.

La trace du radio-altimètre n'indique guère d'écart supérieur à 30 pieds, ce qui paraît justifier qu'en première analyse on ne tienne pas compte des effets dus à de légères variations de l'altitude de l'avion dans le tracé des cartes de la réponse INPUT. De fait, sur l'ensemble des enregistrements, les variations d'amplitudes liées à des sautes locales d'altitude excèdent rarement 20 p. 100.

Choix des paramètres. Pour représenter les résultats et mettre en évidence les variations latérales de la conductibilité des terrains, on a d'abord pensé à tracer une carte en courbes de niveau des variations de l'amplitude de la réponse enregistrée sur le canal 6 (figure 6), puis, afin de mieux préciser les conditions superficielles, on a tracé celle de la réponse enregistrée sur le canal 1 (figure 7). Les deux cartes obtenues paraissent cohérentes avec ce que l'on savait des résistivités des terrains survolés; en première analyse, les anomalies de forte amplitude correspondent aux zones de conductance horizontale élevée (zones à eau salée), tandis que les anomalies d'amplitude réduite correspondent aux zones à conductance horizontale faible (zones à eau douce).

Tout ceci est assez cohérent avec ce que l'on peut attendre de la réponse d'un système électromagnétique à impulsions. Sans entrer dans une discussion théorique difficile (Loeb, inédit) on sait que, si la réponse à une couche superficielle très peu conductrice est généralement faible ou nulle sur les canaux correspondant aux temps d'échantillonnage les plus longs, une réponse forte sur les mêmes canaux peut traduire aussi bien l'existence d'une cause conductrice profonde que celle d'une couche superficielle très conductrice. Nous avons retenu comme un guide très approximatif et souvent ambigu, que la carte correspondant au canal 6 reflète davantage les variations liées à des causes profondes que celle du canal 1.

Ces considérations théoriques jointes à la constatation de divergences entre les deux cartes d'amplitudes nous ont conduit, pour mettre mieux en évidence ces divergences, à établir des cartes des valeurs de rapports des amplitudes enregistrées sur différents canaux. La carte du rapport des amplitudes enregistrées sur les canaux 1 et 6 (V_t=300/V_t=1900) est présentée sur la figure 8. Cette carte est également cohérente, mais sensiblement moins contrastée et détaillée que les cartes en courbes d'égales amplitudes.

Analyse des cartes d'amplitudes (figures 6 et 7). Lorsque l'on procède à l'analyse d'une carte d'égales amplitudes de la réponse électromagnétique transitoire enregistrée à un temps d'échantillonnage donné, on effectue une démarche analogue à celle qui consiste à interpréter la carte d'égale conductivité apparente que l'on obtiendrait d'une prospection électrique au sol avec une longueur de ligne donnée. L'interprétation d'un tel document comporte toujours des ambiguïtés liées à la superposition d'effets causés par des variations de résistivité dans les différentes couches. Dans les cas simples toutefois certaines anomalies sont d'une interprétation plus évidente.

Examen de la figure 6. C'est ainsi que sur la carte d'amplitudes de la réponse sur le canal 6 l'étroite anomalie résistante (amplitude faible), qui coïncide exactement avec le cours du Petit Rhône, peut être interprétée comme l'effet de l'étroit chenal d'eau douce (plus résistante) cheminant à travers une vaste zone (argileuse ou salée) très conductrice. Par analogie, la vaste zone résistante

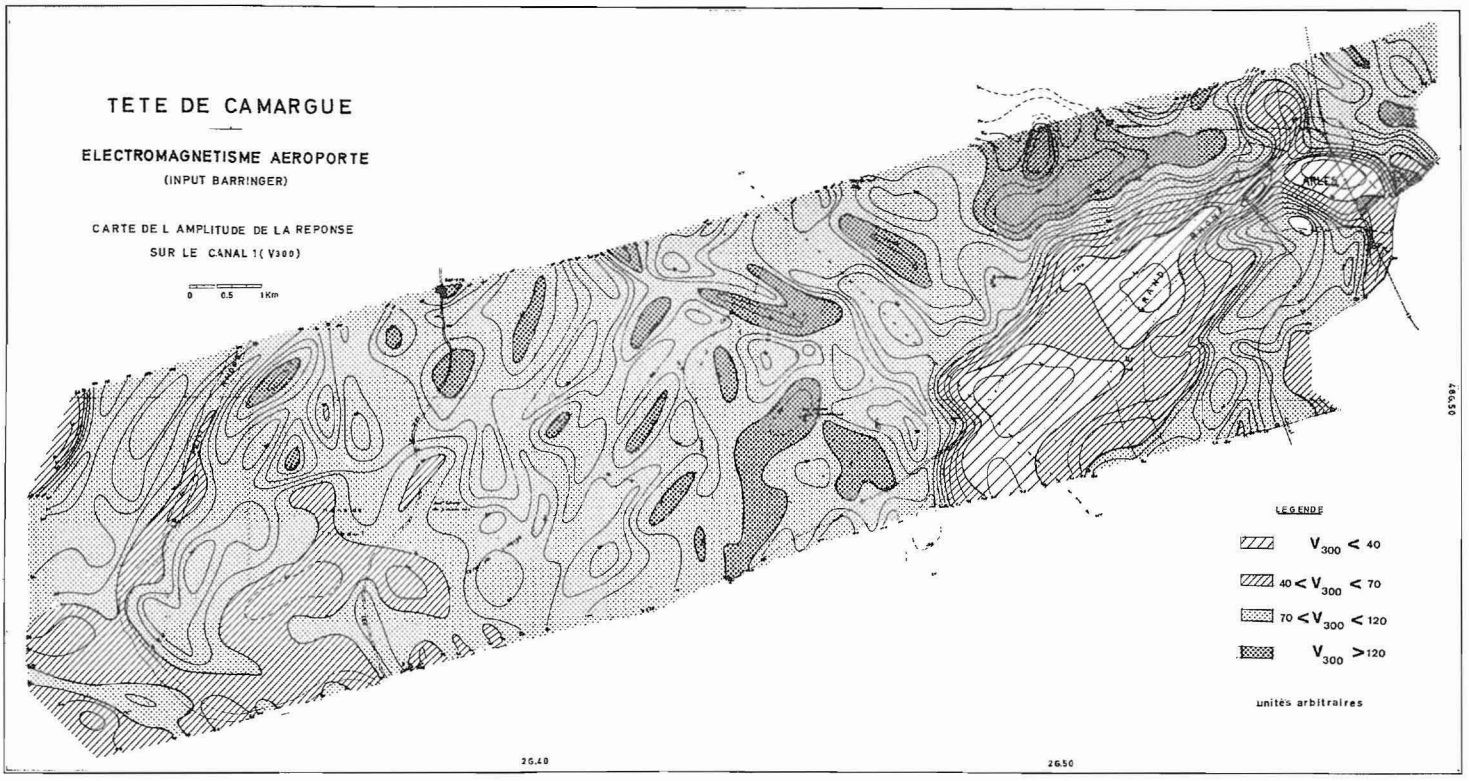


Figure 7. Amplitude de la réponse INPUT sur le canal 1.

s'étendant à cheval sur le cours du Grand Rhône et débordant celui-ci vers l'Ouest de plusieurs kilomètres, peut s'interpréter comme une large nappe d'eau douce alimentée par le Grand Rhône et les cailloutis pliocènes qui affleurent sur la rive gauche.

Il faut noter cependant l'existence dans cette zone d'un axe transversal plus conducteur parallèle à des directions structurales connues (WSW-ENE) et de deux petites anomalies fermées plus résistantes, dont l'une, située au droit de la cité d'Arles, englobe l'affleurement de calcaires jurassiques. Il semble donc bien que dans le voisinage du Grand Rhône l'influence des terrains du substratum se fasse sentir et vienne interférer avec celle des terrains aquifères.

En ce qui concerne le contact entre les eaux douces et les eaux salées, le très fort gradient observé à 2.5 kilomètres à l'ouest de la boucle du Grand Rhône paraît correspondre à une interface très franche. En remontant vers le point de séparation les deux bras du Rhône, le gradient d'amplitude devient moins accentué et son tracé plus sinueux par suite soit de l'étalement de la zone de transition entre eaux douces et eaux salées soit de l'interférence d'autres facteurs d'origine plus profonde ou nettement superficielle. Cette zone de gradient (en particulier la courbe 8 divisions arbitraires) suit de près le contact eau douce - eau salée dans les limons marqués par la courbe d'isoconductance $C_1 = 5$ ohm^{-1}.

Dans la zone située juste à l'ouest de cette limite, il y a coexistence d'anomalies de faible amplitude et de forte amplitude. Il n'est pas possible de dire, au seul vu de cette carte, s'il s'agit de variations locales de la salinité de la nappe (lentilles d'eau douce flottant sur l'eau salée ou de phénomènes d'origine différente.

Toute la moitié ouest de la zone prospectée, Petit Rhône excepté, correspond à une vaste zone conductrice à très conductrice, dont les anomalies d'amplitude maximum sont très fréquemment liées aux zones couvertes de marécages. Là encore il est impossible de dire si les différents gradients observés sont significatifs et encore moins d'en apprécier l'origine. Une exception cependant concerne une anomalie allongée et relativement plus résistante qui suit exactement, dans la partie centrale de la zone, le tracé d'un des plus anciens canaux de drainage des eaux superficielles (canal dit du Pont de la Rousty). On est tenté de voir dans cette anomalie l'expression d'un assainissement localisé des terrains superficiels dû à une circulation plus rapide des eaux. D'autres anomalies modulant le fond très conducteur de toute la partie ouest de la zone étudiée peuvent avoir une telle origine superficielle, les divagations du Rhône au cours des périodes historiques ayant pu créer des zones où les eaux superficielles circulent à des vitesses très différentes.

L'intérêt suscité par l'analyse de la carte d'amplitudes correspondant au temps d'échantillonnage le plus long (1900 microsecondes) a conduit à tracer celle qui correspond au temps d'échantillonnage le plus court (300 microsecondes), présentée sur la figure 7.

Examen de la figure 7. Sur la carte d'amplitudes correspondant au canal 1, on retrouve bien certains traits essentiels de la carte d'amplitudes correspondant au canal 6 (figure 6), mais on constate aussi plusieurs divergences notables. C'est ainsi que si l'on retrouve bien l'anomalie étroite et résistante du Petit Rhône, on constate aussi que des anomalies de même amplitude s'étalent de part et d'autre de celui-ci. On retrouve de même la zone résistante du Grand Rhône, mais son extension est nettement plus réduite

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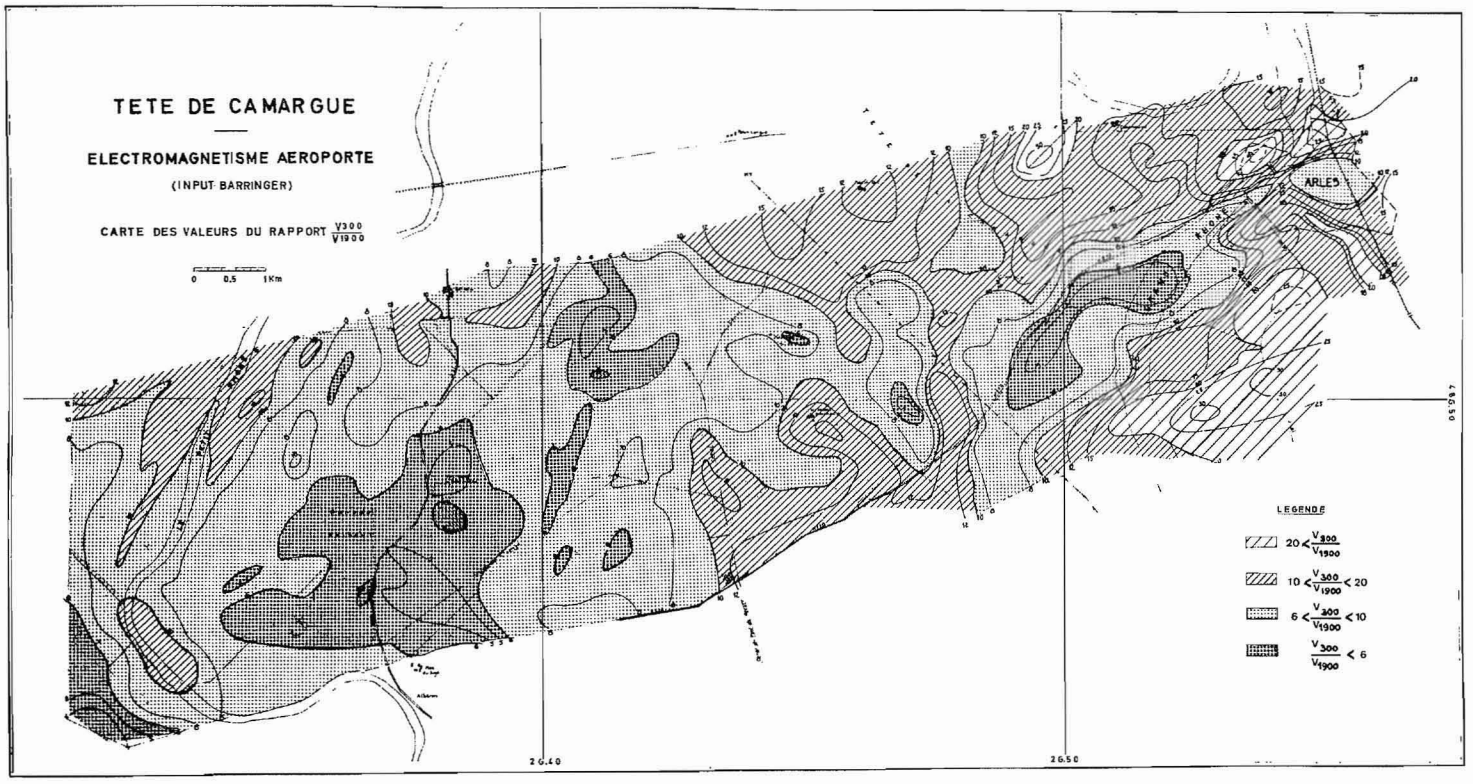


Figure 8. Rapport des amplitudes mesurées sur le canal 6 et sur le canal 1.

sur la carte du canal 1, et si le tracé de l'interface eau douce - eau salée paraît confirmé, on observe par contre trois sortes de divergences:

La disparition sur la carte du canal 1 de l'axe conducteur WSW-ENE paraît confirmer l'origine profonde de cette anomalie.

L'apparition, sur la rive gauche du Grand Rhône, d'une zone conductrice qui n'était qu'à peine suggérée sur la carte du canal 6, paraît correspondre à la présence d'un conducteur à la fois mince et superficiel, puisqu'elle n'affecte que le début de la courbe de décroissance INPUT.

De même le renforcement sur la carte du canal 1 de l'amplitude d'une anomalie conductrice déjà signalée entre les deux bras du Rhône, juste à l'ouest d'Arles, ne peut que suggérer une origine superficielle pour ce conducteur.

Enfin, dans la zone très conductrice couvrant la majeure partie de la plaine comprise entre les deux bras du Rhône, on constate une coïncidence relative des anomalies d'amplitude maximum avec toutefois des désaccords sensibles entre les deux cartes quant aux corrélations de détail. En particulier, l'anomalie dite du canal du Pont de la Rousty n'apparaît pas clairement sur la carte du canal 1.

Analyse de la carte de rapports d'amplitudes (figure 8). La carte présentée sur la figure 8 est celle des rapports d'amplitudes des réponses enregistrées sur les canaux 1 et 6. Elle est beaucoup plus difficile à analyser, surtout dans un cas aussi complexe, que les deux cartes d'égales amplitudes présentées sur les figures 6 et 7. Elle est également moins significative et différenciée en ce qui concerne la délimitation des zones pouvant présenter un intérêt du point de vue hydrologique. En particulier, l'anomalie du Petit Rhône n'apparaît pas nettement (voir exemple de courbe de décroissance sur la figure 5); mais surtout on observe dans la zone pourtant résistante du Grand Rhône des rapports tout aussi faibles, sinon plus, que dans les zones conductrices du centre et de l'ouest. Il ne semble pas que les cartes de rapports d'amplitudes (non présentées ici) établies pour les canaux 1 et 3 (300 et 700 microsecondes) et pour les canaux 3 et 6 (700 et 1900 microsecondes) puissent apporter des informations nouvelles.

De telles cartes des rapports d'amplitudes, qui seules, sont très difficiles à interpréter dans le cas de plus de deux terrains, paraissent néanmoins très utiles pour mettre en évidence les divergences souvent très significatives entre les diverses cartes d'égales amplitudes.

Comparaison avec les résultats d'études au sol

Nous avons jusqu'ici confronté les cartes INPUT avec les grands traits de la géologie et de l'hydrogéologie de la zone. Leur confrontation avec les résultats des sondages électriques réalisés, dont l'objet avait été l'étude des seules formations aquifères, outre qu'elle a aidé à définir le tracé des courbes équiconductance et équirésistance de la figure 3, a confirmé que les réponses électromagnétiques ne pouvaient s'expliquer complètement au seul niveau des terrains aquifères. Cela nous a conduit à réinterpréter les sondages électriques en vue de préciser la conductance du troisième terrain. Les résultats de l'interprétation sont donnés, pour les profils *A*, *B*, *C* et *D*, sur les figures 9, 10, 11 et 12, où l'on a également reporté les résultats électromagnétiques.

Seul le profil *D* (figure 10) orienté NE-SW recoupe perpendiculairement une interface eau douce - eau salée. Celle-ci est marquée par un accroissement de 2.5 à plus de 10 ohm^{-1} de la

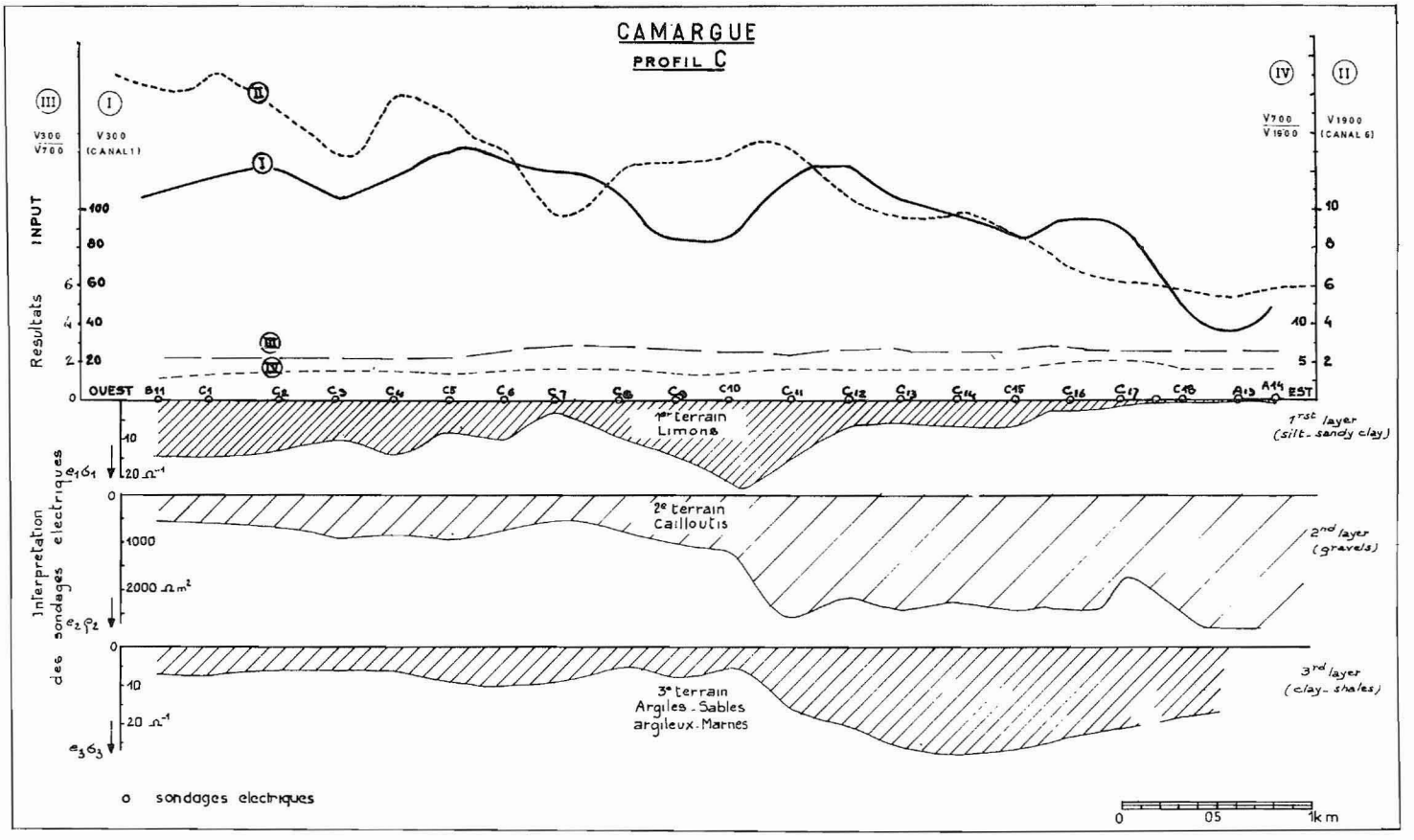
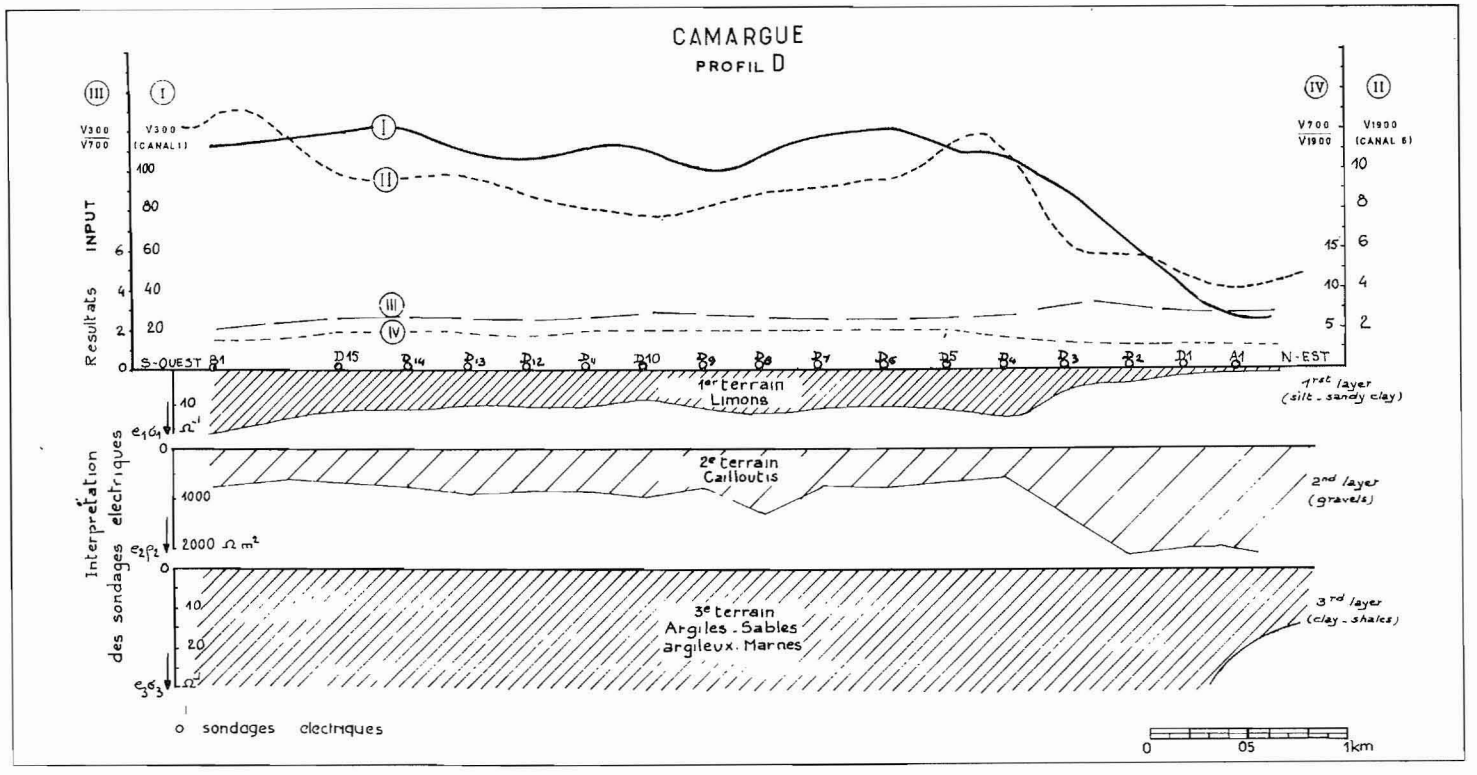


Figure 9. Comparaison des résultats INPUT et des coupes électriques le long du profil *C*.

Figure 10. Comparaison des résultats INPUT et des coupes électriques le long du profil *D*.



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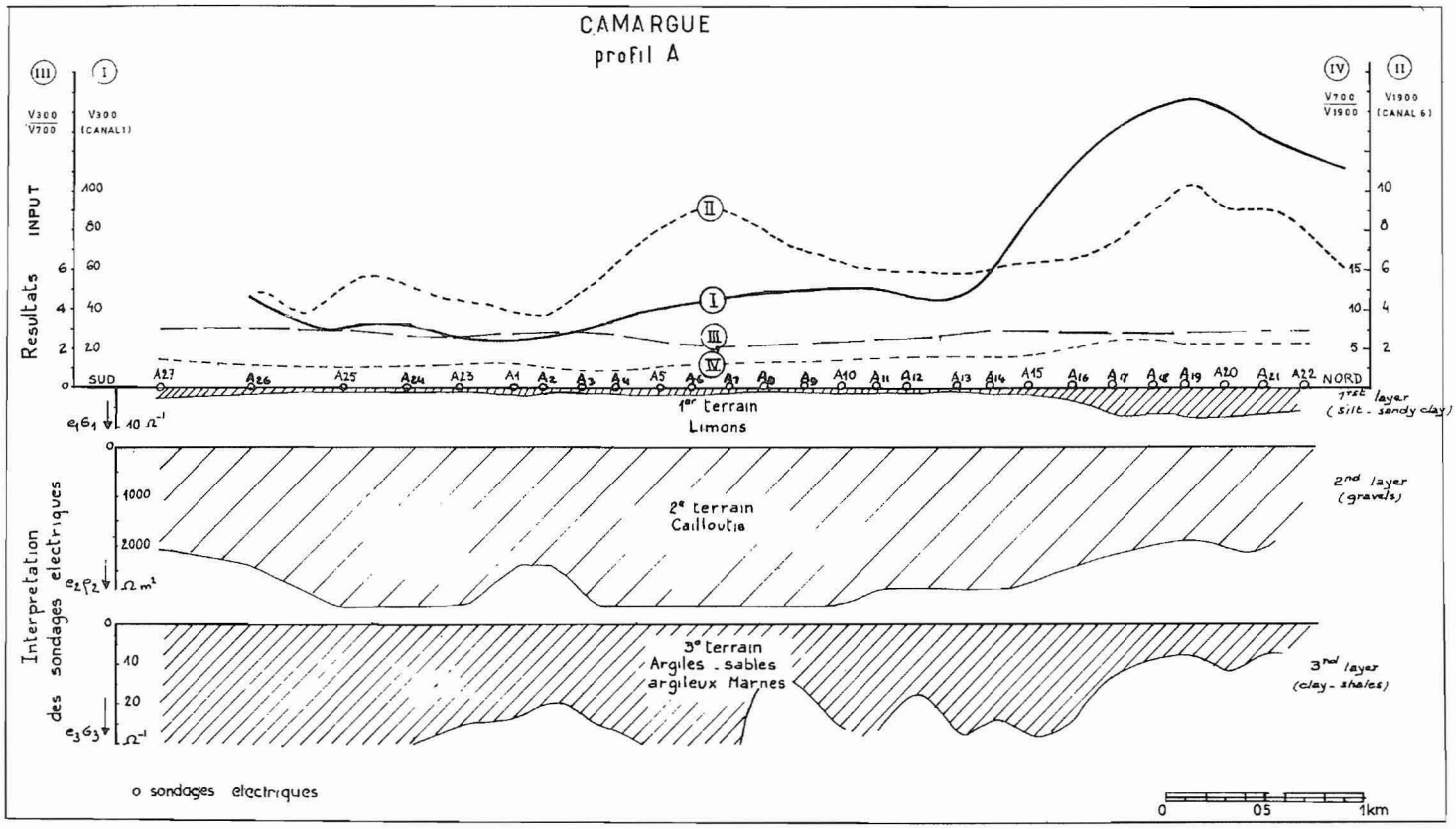


Figure 11. Comparaison des résultats INPUT et des coupes électriques le long du profil *A*.

conductance horizontale du premier terrain et par une chute de 2000 à moins de 700 ohm-m² de la résistance transversale du second terrain. Il y correspond un changement d'amplitude de 6 à 12 unités arbitraires sur la carte du canal 6 (figure 6) et de 40 à 100 unités arbitraires sur la carte du canal 1 (figure 7).

Sur le profil *C* (figure 9) orienté ENE-WSW, la limite vers l'est des limons très conducteurs (argileux ou salés) ne coïncide absolument pas avec l'interface eau douce - eau salée dans les cailloutis aquifères. Le décalage entre les deux limites, bien que supérieur à 1 km, n'apparaît pas nettement sur les profils d'amplitude INPUT. Une remontée des calcaires, marquée par une diminution de 28 à 7 ohm^{-1} de la conductance horizontale du troisième terrain (argiles plaisanciennes), cache complètement la diminution de 2500 à 800 ohm-m² de la résistance transversale du second terrain (cailloutis aquifères) correspondant au passage des eaux douces aux eaux saumâtres.

Sur le profil *A* (figure 11), où la résistance transversale des cailloutis ne tombe jamais en dessous de 2000 ohm-m^2 (zone à eau douce), les sondages électriques permettent de confirmer partiellement deux hypothèses fondées sur l'analyse des écarts entre les cartes d'amplitudes de la réponse EM sur les canaux 6 et 1:

L'axe conducteur transversal observé sur la carte du canal 6, mais invisible sur la carte du canal 1 (*A*6-*A*7 sur le profil), ne correspond à aucune variation au niveau des premier et second terrains, dont la conductance horizontale reste très faible, mais bien à un accroissement considérable mais localisé de l'épaisseur du troisième terrain (enfouissement des calcaires jurassiques probablement lié à un accident tectonique). Cet exemple montre que si les terrains de surface ne sont pas trop conducteurs, la profondeur d'investigation du système INPUT dépasse 100 mètres.

A l'ouest d'Arles, on observe des amplitudes relativement plus fortes sur la carte du canal 1 que sur celle du canal 6 du fait qu'à une augmentation de la conductance des limons se superpose une diminution de la conductance du troisième terrain due à une remontée des calcaires jurassiques (*A*15 à *A*17 sur le profil).

Il y a peu à dire sur le profil *B* (figure 12) où les réponses électromagnétiques paraissent liées uniquement à la conductance horizontale des limons.

En quelques points des profils *B*, *C* et *D* les réponses électromagnétiques restent mal expliquées. Des sautes d'altitude y sont parfois pour quelque chose, mais l'examen des cartes montre qu'il s'agit essentiellement des points où les profils sont parallèles à des zones de transition. Cette circonstance met en lumière le fait que les résultats d'une campagne d'électromagnétisme aéroporté conduirait à mieux implanter les travaux de géophysique au sol. Mais l'analyse précédente montre aussi que cette campagne aéroportée ne suffit pas. L'interprétation ne peut être complète que si des travaux au sol viennent préciser les indications données par les cartes électromagnétiques.

Dans la moitié ouest de la zone prospectée, où l'on ne dispose d'aucun sondage électrique, trois forages pétroliers et plusieurs puits creusés à proximité du bras du Petit Rhône pour des

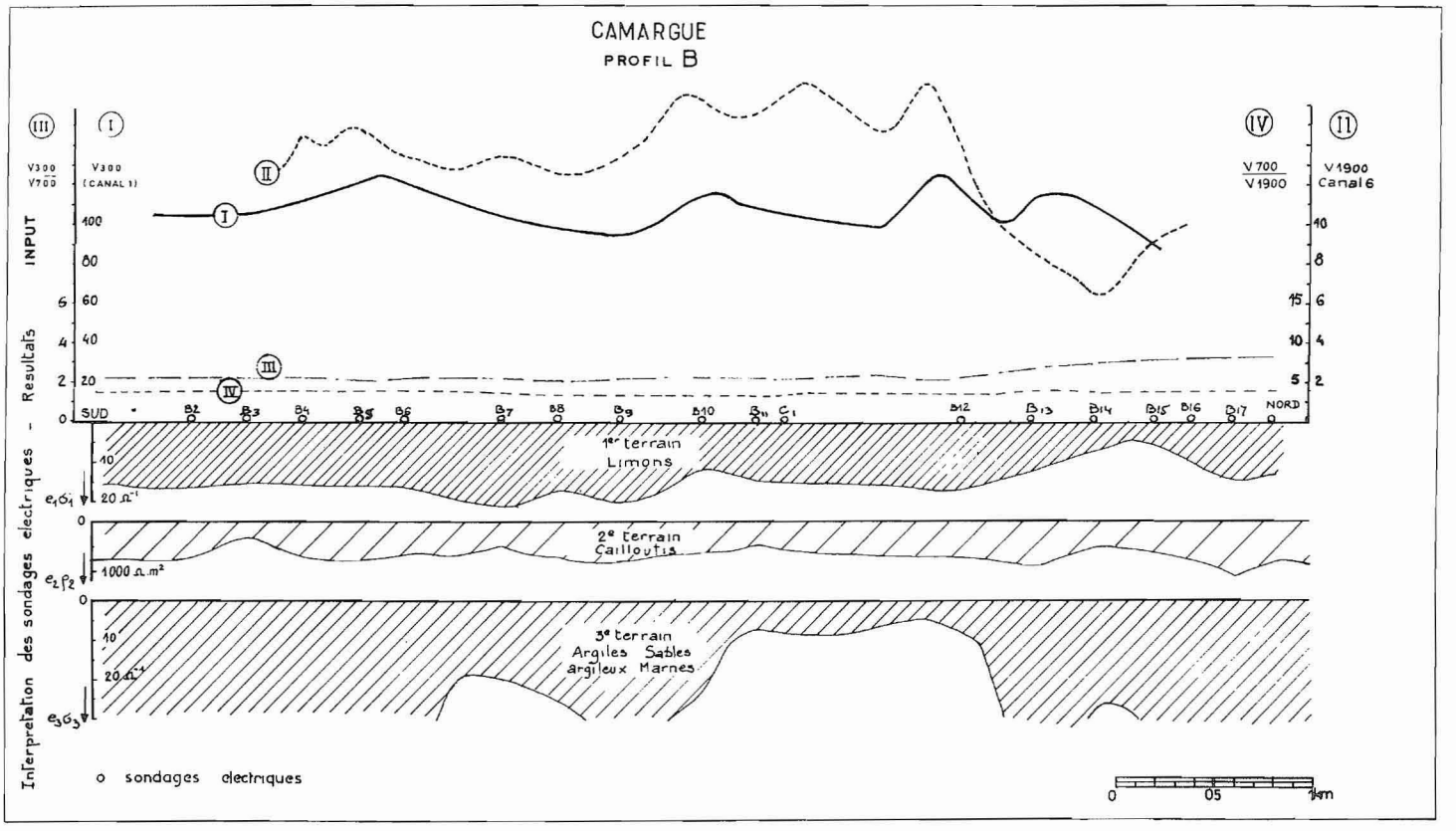


Figure 12. Comparaison des résultats INPUT et des coupes électriques le long du profil *B*.

recherches d'eau montrent que les limons et cailloutis, un peu plus épais, sont envahis par des eaux très chargées en sel, sauf dans une bande étroite le long du Petit Rhône où il existe des captages d'eau douce.

Un forage hydrologique de reconnaissance implanté à quelque 500 mètres du Petit Rhône a donné des eaux salées. Nous ne possédons aucune information sur la hauteur verticale exacte de la tranche d'eau douce correspondant à l'underflow (sous-écoulement) du Petit Rhône.

En résumé, la carte des amplitudes de la réponse sur le canal 1 reflète essentiellement les variations de la conductance horizontale du premier terrain (limons) tandis que sur la carte correspondant au canal 6 l'effet toujours prédominant des variations de la conductance horizontale du premier terrain est modulé par les effets liés à des variations localisées au niveau des couches plus profondes, effets superposés que l'on a pu séparer dans au moins un cas: un changement de la conductance de la troisième couche (variation de profondeur du substratum résistant) est visible lorsque la conductance horizontale des deux premières couches est à la fois assez faible et assez constante. Par contre, le passage des eaux douces aux eaux saumâtres dans les cailloutis aquifères, marqués par un changement de la résistivité du deuxième terrain et qui a lieu sous un premier terrain déjà conducteur, reste peu visible sur la carte du canal 6 même dans les zones où la conductance des premier et troisième terrains varie peu.

Conclusions et recommandations

Le test à caractère hydrologique effectué en Camargue, dans la partie nord du delta du Rhône, montre que la méthode INPUT de prospection électromagnétique aérienne par impulsions peut dégrossir des problèmes, même assez complexes, autrefois réservés à la prospection électrique au sol. Les conditions d'application imposent cependant des spécifications assez strictes:

Il est souhaitable que la *topographie* soit *suffisamment plate* pour éviter autant que possible des calculs de corrections d'altitude lourds et difficiles à maîtriser dans l'état actuel de la technique.

Il est nécessaire de disposer dans les cent premiers mètres de *contrastes de conductance horizontale* au moins aussi favorables que dans les cas traités par prospection électrique au sol.

Il est souhaitable d'effectuer un étalonnage préliminaire et *indispensable d'effectuer des contrôles approfondis par sondages électriques*.

Enfin il faut adapter et coordonner les programmes pour *amortir les frais de mise en place* de l'avion.

En ce qui concerne le choix de la grandeur à cartographier, il semble que les amplitudes des premier et dernier canaux (ou l'une des amplitudes et son rapport à l'autre) soient les paramètres les plus sensibles et les plus maniables du point de vue qualitatif. L'examen comparé des cartes des amplitudes des réponses enregistrées sur les canaux 1 et 6 permet en effet de séparer les variations superficielles de celles qui ont lieu dans les terrains plus profonds et de localiser des zones de transition qu'il reste à contrôler par des travaux au sol.

Les terrains résistants qui, le plus souvent, englobent le niveau aquifère auquel on s'intéresse, ne jouent, en électromagnétisme, que par leur conductance horizontale, laquelle, dans certaines conditions, peut être négligeable devant celle d'autres couches

sans intérêt économique (couverture ou substratum). Dans le cas présenté ici, il ne faut pas oublier que si certains résultats sont spectaculaires, une partie du problème au moins échappe à l'investigation par avion; c'est celle en particulier de la zone de transition dans l'aquifère pourtant intéressante du point de vue hydrologique. En effet, par suite de l'hétérogénéité des terrains et de phénomènes superficiels, la nappe d'eau douce contenue dans les cailloutis peut s'étendre nettement au-delà de la limite d'invasion des limons par les eaux salées. L'étude des zones de transition reste donc bien du ressort des techniques d'exploration au sol.

De façon plus générale, l'origine superficielle ou profonde de telle anomalie peut ne transparaître d'un examen des différences entre diverses cartes électromagnétiques que si les paramètres géoélectriques ne varient pas trop dans les autres couches. Là encore les sondages électriques sont indispensables pour lever le doute. Cependant il faut bien préciser qu'un nombre restreint de sondages électriques implantés judicieusement doit permettre d'élucider les points essentiels soulevés par l'interprétation des cartes INPUT. On peut considérer que dans une région assez homogène, et c'est le cas de la Camargue, un même nombre de sondages électriques, s'ils avaient été implantés après les vols, aurait pu conduire à l'interprétation approfondie d'une zone cinq à dix fois plus vaste.

Dans l'état actuel de la question, l'interprétation des données de la méthode INPUT reste essentiellement qualitative. Des spécifications plus sévères, peu compatibles d'ailleurs avec les conditions de vol à très basse altitude, seront imposées le jour où l'on introduira l'enregistrement numérique avec de nouvelles possibilités de traitement de l'information. Nous n'en sommes pas encore tout à fait là. Mais d'ores et déjà, nous disposons d'un outil permettant d'obtenir rapidement et à une échelle inusitée, des cartes assez élaborées de la conductance longitudinale, même dans des régions d'accès très difficile.

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Discussion on Étude par prospection électromagnétique aérienne d'un contact eau douce - eau salée dans le delta du Rhône by P. Baudoin, G. Durozoy and M. Utard

A. Becker, Geological Survey of Canada. I was very interested in the slide where you show the thickness-conductivity product. Could we have a look at one of those slides again?

P. Baudoin, Bureau de Recherches Géologiques et Minières, France. (Referring to Figure 9, profile *C*) The solid line is the INPUT response on channel 1, the dotted line is the response on channel 6. Below are the interpretations. From top to bottom in this figure are the thickness-conductivity products of the three layers.

M. Utard, Geoterrex CGG, France. The interpretation of the ground resistivity indicates that the upper layer is sandy clay containing fresh water and the second layer is composed of a compact zone with gravels and fresh water, so that there is no conductivity change in the alluvium. We could only explain the bump on channel 6 by a deeper feature. We had to reinterpret all electrical soundings to find what is at depth; we found right in the middle a sharp increase in the conductance of the terrain underneath the alluvium. Geologically such a feature corresponds to an increase in the thickness of clay which is estimated to be at least four times the thickness of clay on either side of the feature. This feature is between 100 and 200 metres below surface and probably related to a fault known to occur farther east. The penetration of the EM field is limited by the conductance of the upper layer. A very conductive layer at the top, for example, a sheet of infinite conductivity, would act as a mask, thus cancelling the primary field by reflecting up with a change of sign. It is very difficult to see any change in the conductivity below a sheet of sea water. It is almost impossible, I would think, except by using some very sophisticated techniques such as one used by Dr. Barringer in Sweden over sea water. He could see a conductor underneath the sea water.

Baudoin. The conductance of the first layer was about 1 mho, the second about 5 - 6, and the third layer was more than 10 - 15 mho. Conductance is the thickness divided by the resistivity. Depths for the two first layers is almost constant, 20 metres for the first and 40 metres for the third. So what we see is the change in resistivity.

Becker. Thank you very much indeed. I think that this is a very excellent example.

Utard. If you try to use INPUT as a reconnaissance tool for mapping superficial geology compared to other classic methods on the ground, resistivity mapping with only one constant electrode array, you map what happens at an average depth. But by analysing the INPUT transient at various times after the decay, you can draw up several horizontal sections. You can recognize the superficial changes and you can detect a resistivity change deeper down. Of course, you have some ambiguity which has to be checked by electrical sounding or other classic methods. Resistive layers are not seen by the INPUT method. You need a rather good contrast of conductivity to obtain a fair change in the transient response. The requirements for applying this method are even more severe than for the classic resistivity method. If you look at the problem of how to integrate this type of airborne investigation into a general scheme, you must know in advance the ground conditions prior to the commencement of a systematic coverage.

R. Giret, Compagnie générale de Géophysique. This survey in the Rhône

valley entailed 120 line miles. It was the first time that we conducted an INPUT survey. We did not apply any mathematical method of interpretation. The profile that was presented has only shown what can be done and information obtained. But we are far from a normal user's method. I think if we use 6 or 12 or more channels with digital recording and fly thousands of line miles of survey, we have to use digital methods to process the data. With the parallel development of the method of interpretation of digital recorded data and of flying, more test surveys will probably give within one or two years a lot more information on the method. I think that it is a very rich method as Mr. Utard has explained several times. I think that the tests that have been prepared by the Geological Survey of Canada are very important since the data will be able to be handled in a more sophisticated fashion.

L.S. Collett, Geological Survey of Canada. I couldn't agree more on this point. We are just starting. There is a lot more information can be had by putting a second coil, and even a third coil, in this aircraft.

La prospection géophysique et la recherche des eaux souterraines

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Paris, France

Résumé. La présente étude traite tout d'abord des principales méthodes employées pour la prospection des eaux souterraines, et aborde ensuite certains types de problèmes hydrogéologiques qui peuvent être résolus par des techniques de géophysique.

Les principales méthodes électriques comprennent: la résistivité, l'enregistrement du profil, la cartographie de la résistivité et le sondage électrique. Les méthodes électriques offrent de nombreux avantages. Généralement, la réfraction séismique est utilisée pour suppléer aux méthodes électriques.

Les méthodes de polarisation spontanée sont d'un usage plutôt restreint, et un exemple en est donné. Les méthodes de polarisation provoquée sont encore extrêmement controversées et ne sont que rarement employées. Les méthodes magnétiques peuvent être employées dans des cas vraiment spéciaux, par exemple dans certains complexes volcaniques, tandis que les méthodes électromagnétiques donnent des résultats douteux et sont d'un usage restreint.

Les principaux problèmes hydrogéologiques traités par la géophysique comprennent les formations alluvionnaires, les nappes phréatiques perchées, la recherche des bassins, les aquifères des dunes, et les études de reconnaissance des aquifères pénétrées par l'eau salée. Dans le domaine des études structurales, la géophysique peut servir à la démarcation des horsts, des fossés tectoniques, des filons-couches aquifères, des arènes granitiques et des zones de fracture. La circulation des eaux karstiques est un problème qui ne peut être que rarement résolu par la géophysique.

Les méthodes de diagraphie électrique sont d'un emploi généralisé. Légers et économiques, les instruments électriques utilisés en géophysique sont également ceux que l'on emploie le plus en hydrogéologie.

Abstract. The paper deals first with the principal methods which are applied to groundwater investigations and then with types of hydrogeological problems to which geophysical techniques can be applied.

The principal electrical methods are: resistivity, profiling, resistivity mapping and electrical sounding. Electrical methods have many advantages. Seismic refraction is generally used to complement electrical methods.

Natural polarization methods have a somewhat restricted use and an example is given. The induced polarization methods are still very controversial and are used very little. Magnetics can be used in very special cases, as in some volcanic complexes, while electromagnetics give uncertain results and have restricted application.

The main hydrogeological problems dealt with by geophysics are alluvial formations, perched water-tables, the search for basins, sand-dune aquifers and the reconnaissance studies of aquifers penetrated by saline waters. For structural studies, geophysics can be used for delineating horsts, grabens, water sills, granitic sands and fracture zones. The circulation of karstic waters is a problem which can rarely be treated by geophysics.

Electrical logging methods have wide application. Still the most widely used geophysical methods in hydrogeology are the lightweight and economical electrical methods.

Les deux principales méthodes de prospection géophysique appliquées à la recherche des eaux souterraines sont la prospection électrique (méthode dite des résistivités) et la réfraction séismique, la première étant de beaucoup la plus utilisée, car elle permet de traiter le plus grand nombre de problèmes, rapidement et économiquement.

La prospection électrique par résistivité comprend deux techniques principales: le sondage électrique et le profil de résistivité, donnant lieu à la carte de résistivités. Parmi les autres méthodes, on peut citer: la polarisation spontanée, la polarisation provoquée, l'électro-magnétisme et le magnétisme; ces méthodes peuvent être appliquées en général à des problèmes particuliers.

En général, il faut admettre qu'il n'y a pas en hydrologie de détermination directe du minéral recherché, ici l'eau.

L'application de la géophysique consiste à déterminer l'allure des couches, leur faciès et, en quelques cas particuliers, leur perméabilité (cas des couches alluviales), ou mieux les zones où la perméabilité est la plus élevée sur une zone donnée.

On se ramène ainsi au point de vue de la géophysique pure à la recherche des hydrocarbures et jusqu'à un certain point à celle des minéraux métalliques. Pour les hydrocarbures, la géophysique détermine des structures favorables sans plus. Pour les gîtes métalliques, elle permet d'aller un peu plus loin puisque souvent elle met en évidence des concentrations de minéraux présentant une certaine conductibilité ou des substances disséminées considérées comme électriquement résistantes par les autres méthodes (cas de la polarisation provoquée).

Le cas des nappes alluviales mérite une attention spéciale, car on frise cependant la recherche directe. En effet, la géophysique permet de déterminer s'il existe un ancien lit et dans cet ancien lit si son comblement comprend des formations perméables, ou des zones de plus grande perméabilité. S'il y a de l'eau, c'est évidemment le long de cet ancien lit qu'on la trouvera et souvent en quantité, si bien que tout puits ou forage bien placé sera productif presque à coup sûr. On voit ainsi que dans ce cas on arrive presque à la recherche hydrologique directe.

Étendons-nous un peu sur la notion de perméabilité. Pour les alluvions et formations meubles, j'ai dit que la résistivité permettait de comparer, qualitativement tout au moins, la perméabilité des différentes zones étudiées. C'est ainsi que pour une même eau d'imbibition, une alluvion sera d'autant plus perméable que sa résistivité sera plus élevée. Si l'on est en présence d'une roche compacte, une zone faillée ou une zone de fractures coïncidera avec une bande relativement conductrice sur une carte de résistivité.

La méthode de la carte de résistivité consiste à mesurer la résistivité d'une tranche de terrains d'épaisseur constante et de telles cartes peuvent être établies pour plusieurs profondeurs d'investigation afin de mettre en évidence différents épisodes stratigraphiques.

Le sondage électrique est la technique la plus complète de la prospection électrique puisqu'elle permet de mesurer d'une façon continue la variation de la résistivité du sous-sol pour des profondeurs allant en croissant et dont on est le maître.

Cette méthode permet ainsi de déterminer facilement les zones à eau salée et leur contact avec les zones à eau douce. On sait, en effet, que la résistivité d'une formation est d'autant plus faible que son eau d'imbibition est plus salée (la résistivité d'un sable à eau de mer est de 1.2 ohm–m environ, alors que celle d'un sable à eau douce peut atteindre plus d'une centaine d'ohms–m).

J'ajouterai que la prospection électrique est d'une grande facilité de mise en oeuvre sur le terrain, qu'elle est très souple et immédiatement adaptable à bien des problèmes différents; en outre, son application est économique (la Conférence inter-africaine d'hydrologie, qui s'est tenue à Nairobi en janvier 1961, a chiffré de 50 à 70 p. 100 l'économie réalisée sur les forages implantés avec l'aide de cette prospection).

Certes, elle ne peut résoudre tous les problèmes et la condition essentielle pour qu'il y ait succès est qu'il existe une certaine continuité géologique dans le sous-sol, ce qui, par exemple, ne lui permet pas de traiter certains problèmes dans des formations volcaniques et les terrains métamorphiques redressés et très plissés.

Il existe plusieurs techniques de mise en oeuvre du sondage électrique. Il va de soi que seules sont à retenir, celles qui sont basées sur une interprétation mathématique des courbes expérimentales des sondages électriques et non sur des méthodes empiriques. Cette discrimination circonscrit le choix entre la technique ≪Schlumberger≫, qui mesure le champ électrique au centre du quadripôle utilisé et la technique ≪Wenner≫ qui mesure la différence de potentiel entre deux électrodes à écartement variable et toujours maintenu égal au tiers de la longueur de la ligne d'envoi de courant, façon de procéder qui provoque l'apparition de phénomènes perturbateurs appelés ≪à-coups≫ de prises qui rendent l'interprétation délicate et en limitent la précision.

La seconde méthode utilisée est la *réfraction séismique à faible profondeur.* Au point de vue de la détermination de l'allure structurale du sous-sol, la méthode se rapproche de la technique du sondage électrique puisqu'elles font toutes deux appel à des trajets d'énergie subhorizontaux.

La réfraction séismique est principalement employée pour étudier l'allure du rocher sain. Comparée à la prospection électrique, elle est moins sensible aux faciès des terrains et ne fournit que peu d'indications sur la perméabilité. D'autre part, sa mise en oeuvre est plus complexe, les calculs d'interprétation plus longs et il y a les servitudes du transport, du gardiennage et de l'utilisation des explosifs. Ce qui fait que son prix de revient est plus élevé que celui du sondage électrique.

En ce qui concerne la recherche hydrologique, il me semble que la réfraction séismique soit plutôt une méthode d'appoint à la prospection électrique, sauf dans des cas particuliers (étude d'une nappe alluviale à substratum rocheux résistant, recherche des zones de fractures).

Parmi les méthodes de prospection électrique, on a songé à utiliser les courants naturels de *polarisation spontanée,* phénomènes classés comme parasites en recherche minière et provoqués par l'électro-filtration et l'évaporation et qu'on enregistre à la partie superficielle des terrains détritiques et de certaines roches poreuses. Je ne pense pas que l'on puisse suivre directement avec cette méthode; l'écoulement de nappes phréatiques; mais utilisée judicieusement, en reconnaissance générale, la polarisation spontanée peut permettre de localiser des zones de plus grand coefficient d'infiltration de l'impluvium et même des nappes peu profondes, comme cela a été le cas par exemple en Côte d'Ivoire et au Dahomey. Je pense que cette expérimentation serait intéressante à poursuivre en zones tropicales et semi-arides.

Il y a, par ailleurs, la *polarisation provoquée* (Induced Polarization), méthode qui visait à la recherche directe de l'eau dans les sédiments meubles. Il faut reconnaître que les essais se sont montrés tout à fait décevants, les phénomènes enregistrés étant principalement fonction de la teneur en argile, si bien que son application est à discuter, sinon à déconseiller.

Je citerai maintenant pour mémoire la *méthode magnétique* dont l'utilisation n'a aucune portée générale dans le cas qui nous occupe, mais à laquelle on peut néanmoins faire appel pour certains problèmes concernant les formations volcaniques basiques, par exemple pour repérer des coulées de lave basaltique ou des dykes de basalte ayant un rôle hydrologique.

Je terminerai cette liste en mentionnant *l'électromagnétisme.* On sait que cette méthode, susceptible d'une application aéroportée, est essentiellement utilisée pour reconnaître les minerais conducteurs. Certains expérimentateurs ont songé à l'utiliser pour repérer des nappes très superficielles en milieu résistant (par exemple dans l'Ouest africain). J'ajouterai que les meilleures indications ont été enregistrées sur les Chotts salés, ce qui n'a rien d'étonnant.

Nous arrivons maintenant à l'énoncé des *principaux problèmes hydrologiques* qui sont couramment traités par la géophysique. En premier lieu viennent les études de nappes alluviales comprenant l'estimation des épaisseurs d'alluvions, la détermination des anciens lits et sur leur tracé, l'indication des zones où les alluvions présentent la meilleure perméabilité et la détermination de la nature de substratum.

À ces études se rattachent les problèmes de nappes perchées, sur les plateaux ou les terrasses fluviales, ainsi que la recherche des cuvettes dans les régions où le plancher est imperméable.

Il y a les études de nappes dunaires au voisinage de la mer ou des sédiments côtiers, prospections qui consistent à établir une carte en courbes de niveau du substratum et à estimer l'étendue et l'épaisseur de la formation imbibée. Un point important que j'ai déjà signalé précédemment, est la mise en évidence des zones envahies par l'eau salée et de celles comportant une nappe d'eau douce reposant sur l'eau salée, ainsi que les secteurs uniquement à eau douce.

Il s'agit par ailleurs d'études tectoniques pour préciser en profondeur l'allure d'un terrain connu en affleurement et susceptible d'être aquifère (calcaires, grès) et pouvant en outre faire partie d'un système artésien. Ces études conduisent à la détermination de failles, horsts et grabens, anticlinaux et surtout à la recherche de seuils hydrauliques.

On pourra aussi être en présence de formations volcaniques dont il conviendra de mettre en évidence les coulées de laves et les tufs perméables au milieu des tufs argileux et cinérites, ou encore déterminer l'ancien réseau hydrographique sous ces formations.

On a également traité le problème de la recherche de points d'eau dans les arènes granitiques de faciès généralement très variable. Puis celui de la détermination des fractures, zones de broyage et failles affectant un socle cristallin ou cristallophyllien et qui jouent le rôle de drains collecteurs. À ce problème se

rattache l'étude de l'altération superficielle des terrains schisteux plus ou moins métamorphiques et qui peut être liée à certaines réserves d'eau. Entrent aussi dans ces études les problèmes relatifs aux eaux thermales et qui peuvent se résumer dans la recherche d'extensions de failles ou de nouveaux griffons.

Mais un des problèmes fondamentaux reste l'exploration hydrologique des bassins considérés chacun comme une entité, de façon à évaluer les eaux souterraines et contribuer ainsi à l'établissement du bilan hydrique.

En résumé, peu de problèmes échappent en fait à une étude géophysique, à part toutefois certains d'entre eux que pose le véritable régime karstique des eaux dans les calcaires.

Le forage est un ouvrage coûteux dont il convient de tirer le plus d'informations possibles. Pour cela, une opération de *carottage électrique* devra être effectuée à la fin de la perforation et avant la pose du tubage. Ce carottage électrique a un double but: tout d'abord, un but technique en donnant des renseignements cruciaux sur la situation et l'épaisseur des différentes nappes aquifères (principalement dans les complexes argilo-sableux); ensuite, en mesurant la résistivité des différents horizons traversés, il permettra le cas échéant, de reprendre par la suite l'interprétation des sondages électriques et augmenter la précision des estimations de profondeurs.

Conclusion

Ainsi que pour les études pétrolières où l'intervention de la géophysique n'est plus discutée depuis longtemps, on constate que l'on a de plus en plus recours à elle pour les recherches hydrologiques bien conduites. Certes, les études pétrolières permettent l'utilisation de méthodes coûteuses, il n'en est pas de même pour l'eau dont les investissements mis en jeu et leur rentabilité ne sauraient absolument pas être comparés. Aussi convient-il d'adopter des méthodes géophysiques légères et bien éprouvées, d'une application aussi générale que possible et bon marché, et je pense en disant cela avant tout à la prospection électrique qui réunit tous ces caractères.

Geophysical studies in permafrost regions in the U.S.S.R.

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Abstract. Nearly half the territory of the Soviet Union is covered by permafrost. It is widely studied by geophysics especially by geoelectric methods.

The resistivity of frozen soils depends on the ratio of ice and water contained in the pores. Sand resistivity increases sharply when temperature drops below −0.5°C. Clay acquires high resistivity with a drop in temperature to from −2 to −6°C. The velocity of propagation of elastic waves considerably increases in frozen soils. Hence in sections of homogeneously frozen soils, some characteristic horizons with different resistivities are marked out.

In cases of low-temperature permafrost resistivity of Quaternary deposits exceeds the resistivity of bedrock. This makes it possible to determine the depth of the lower interface of the overburden using the decrease of resistivity in electrical soundings.

Some difficulties arise due to big value of ratio ρ_2/ρ_1 and inconsistency of ρ_2. A special procedure of quantitative interpretations of the data is suggested.

By electrical soundings the upper and lower surfaces of permafrost can be determined. Electrical profiling, some other geoelectrical prospecting and thermal methods are applied to map it.

Measurement of rock resistivity conducted in fixed-time intervals make it possible to determine shifts in the permafrost boundary.

Observations on the advance and retreat of permafrost are important for populated areas, mines, hydrotechnical and other engineering projects.

Résumé. Près d'une moitié du territoire de l'Union Soviétique est atteinte par la congélation perpétuelle. Pour son étude on utilise largement les méthodes géophysiques et, en premier lieu, géoélectriques.

La résistivité de la roche gelée est définie par le rapport de la glace et de l'eau, se trouvant dans ses pores. La résistance des sables monte brusquement à la baisse de la température au-dessous de −0.5°C. Les roches argileuses prendent une résistance élevée à la baisse de températures de −2° jusqu'à −6°C. Dans les roches gelées on observe aussi un accroissement appréciable des vitesses de la propagation des oscillations élastiques. Dans la coupe de roches homogènes gelées on peut délimiter un nombre d'horizons électriques caractéristiques, se distinguant considérablement par leur résistance.

La résistance des diluviens en condition de congélation à baisse température se présente ordinairement plus élevée que celle des roches-mère. Cela permet de définir la profondeur du gisement de la frontière inférieure du diluvien suivant la baisse de résistances sur les courbes de sondage. Pourtant cela donne naissance à une série de difficultés liées au rapport élevé ρ_2/ρ_1 et l'inconstance de ρ_2. On a proposé des procédés spéciaux de l'interprétation quantitative.

Les sondages électriques permettent en outre de définir les frontières supérieure et inférieure de la congélation. Pour l'établissement des cartes de la congélation on utilise le profilement par la méthode de résistances, ainsi que quelques autres moyens de prospections géoélectrique et géothermique.

Permafrost covers half of the U.S.S.R. as well as 24% of the continental area of the globe. In these areas, the temperature of the soil and rock is below zero and hence they contain ice in their pores. The thickness of this frozen layer may vary from several meters to hundreds of meters depending on the regions.

Two types of permafrost can be considered according to the temperature: the high-temperature range (0° to −2° or −3°C) and the low-temperature range (down to −8° or −10°C). Depending on the latitude, local thermal conditions, type of soils, vegetation, electrolyte in solution and other conditions, the thickness of the active layer varies from 0.2 to 0.4 m to more than 10 m. Inside the permafrost itself there are often unfrozen portions of various sizes and configurations. There may, for instance, be a stratified distribution or unfrozen section running completely through a thick frozen layer.

Geological mapping of permafrost presents great difficulties. The cost may be ten times higher than elsewhere. That is why large-scale geophysical investigations have been started especially after the commencement of heavy industrial development of northern Siberia. These methods have been used for structural geology and prospecting for water, gold, platinum and other minerals. The permafrost is studied too, as a physical phenomenon in the geological environment.

Some physical properties of frozen soils

When water freezes the properties of rocks are modified. The electrical resistivity as well as the velocity of elastic waves increase as a larger percentage of water is transformed into ice. This together with the specific temperature conditions lead to the definition of permafrost as a particular physical medium.

For temperatures above freezing, the ionic conductivity of soils depends on the mineral composition, the amount of water and air. For frozen soil, you must include ice as an additional component. The ratio between ice and water depends on temperature variations, porosity, dimensions of the pores, concentration of salt solutions and the mineral composition of the soil. The resistivity of pure ice, on the other hand, approaches that of the rock minerals.

The resistivity of the frozen rock is then defined by the ratio between the two phases of water. The glaciation diagrams, that is the amount of ice contained in the soil versus temperature, are closely related to the resistivity versus temperature diagrams. The resistivity of unfrozen soil depends on the total amount of its water content. When the soils are frozen, it depends mainly on the ratio between free water and connate water, the latter freezing at temperature much lower than the former. The resistivity of rocks containing a large amount of connate water increases much more slowly with decreasing temperature than soils with free water. Figure 1 shows the resistivity ρ versus temperature t diagrams for clay and saturated sand.

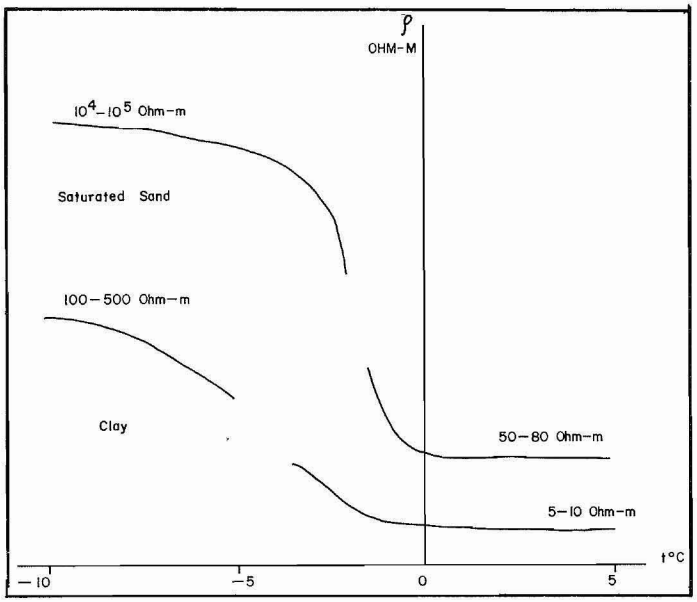


Figure 1. Resistivity versus temperature for saturated sand and clay.

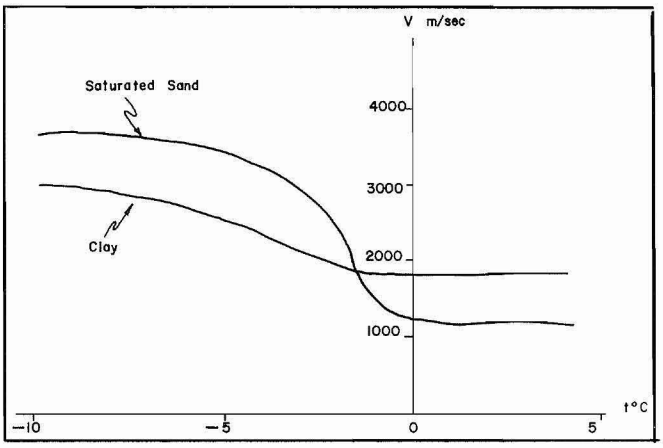


Figure 2. Velocity of the longitudinal waves versus temperature for saturated sand and clay.

For coarse sand and gravel the resistivity increases as soon as the temperature drops below freezing. Finer sand grains produce a smaller rate of increase of resistivity. Clay, on the other hand, shows higher resistivities only when temperatures fall below −2° to −6°C.

As is well known, liquids may show a phenomenon of supercooling. A sand saturated with water may still contain as much as 20 to 30% of unfrozen water after six hours of freezing whereas it will contain only 3 to 4% after three days. Therefore, for a given drop in temperature, the equilibrium resistivity is reached only after a long time. The larger the specific area of the soil, the longer will be the time to reach equilibrium. This temperature hysteresis is sometimes neglected in experiments, thus leading to errors in the measurements.

The dielectric constant of frozen soils has not yet been fully investigated. It is sometimes said to decrease rapidly between 0 and −2°C, as the amount of ice in the soils increases.

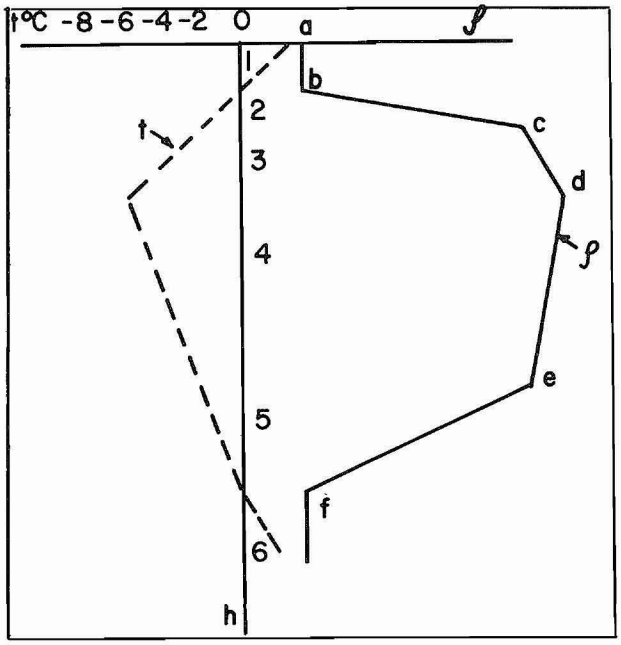


Figure 3. Electrical cross section for soft homogeneous Quaternary sediments in a low-temperature permafrost area (from Borovinsky). ρ – electrical resistivity, t – temperature, 1 to 6 – electrical horizons, h – depth.

Freezing does not seem to change appreciably the polarization properties of soils. This favors the use of the Induced Polarization method for mineral prospecting in permafrost areas which exhibits high resistivity.

Besides the resistivity, the elastic properties of the soils depend on the ratio between water and ice. The increase of the amount of ice just below 0° leads to a sharp rise in the velocity of elastic waves and the modulus E (Figure 2). This rise, however, is not as sharp as that exhibited by the resistivity. It is particularly noticeable in the sand. Because of the inherent state of water in clay, the increase of velocity and of E is shifted towards the lower negative temperature and does not level off.

For saturated sand, the ratio between frozen soil velocity and unfrozen may reach 4 to 5; for clay, it is never more than 2.

The absorption of elastic waves shows a nonlinear variation with temperature. It seems to be rather complicated and has not yet been studied in detail.

Electrical cross section of permafrost

Electrical cross sections of permafrost may vary widely from one place to the other. They depend on the lithology, porosity, alteration, water content, as well as the degree of freezing, temperature and season. In summer, when most of the geophysical investigations are carried out, six electrical horizons can be recognized in a uniform geological cross section consisting of low-temperature frozen Quaternary sediments as shown by Yakoupov (Figure 3).

The first horizon represents the active layer, the one which thaws in summer. The thickness of this first layer varies from a few centimeters to 6 to 10 m. Its resistivity (*ab* in Figure 3) varies

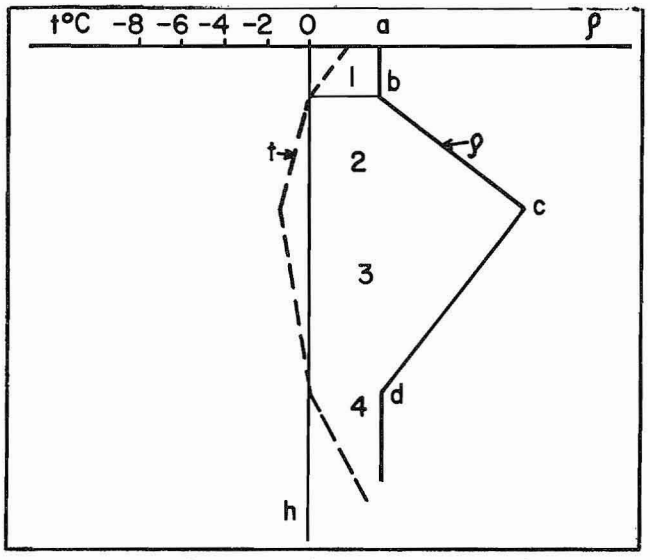


Figure 4. Electrical cross section for a high-temperature permafrost. ρ – electrical resistivity, t – temperature, 1 to 4 – electrical horizons, h – depth.

in the same limits as for unfrozen soils. Depending on their lithologic composition and on the water content, the resistivities will vary from tens to hundreds of ohm-meters. In winter, this layer freezes again completely or partially and the resistivities are increased by ten, a hundred or a thousand times.

The 0°C isotherm constitutes the upper limit of the second layer. Depending on the particle size, the water content and other properties, it extends down to either the −0.5° or −3°C isotherm. For instance, for saturated gravel and sand the lower boundary occurs at −0.5°C. For a compact material the resistivity increases by a factor of a hundred or a thousand (*bc* in Figure 3); for soils with small grain size, the increase is not as great.

In the northeastern part of Siberia, i.e., in a low-temperature permafrost region, this second layer, which extends in this case between the 0°C and the −2°C isotherms, is not usually thicker than the active layer.

The third horizon extends from the lower limit of this previous layer, where the resistivity increased considerably, down to the depth of stable temperature, which in the case of stationary conditions corresponds to the minimal temperature. In this layer, the increase of resistivity follows the temperature variations (*cd* in Figure 3). This increase is in all cases very small. However there is a difference between formations of fine and coarse particle size. The thickness of this layer is not great and it may be compared to the previous ones.

In the fourth horizon, the resistivity drops gradually from this minimal temperature depth to the −3° or −0.5°C isotherm depending on the type of soil. Again, in this case, the lower boundary takes place where a sharp drop in resistivity occurs as above (*de* in Figure 3). The thickness of this layer varies widely and may be about hundreds of meters.

The fifth horizon is similar to the second one. The temperature range is the same and it leads to the same large change in resistivity. This change, however, occurs over a larger thickness, i.e., the thickness between the upper isotherm 0°C to −0.5° or −3°C is smaller than the thickness between the lower isotherms −3° or −0.5°C to 0°C (*ef* in Figure 3). The slope of the change in resistivity with depth is then smaller.

Compared resistivities of frozen and unfrozen bedrock

Type of Rock	Resistivity (ohm-m)	
	Unfrozen	Frozen
Schist	100 - 200	1300 - 1500
Sandstone	200 - 400	3000 - 5000
Granite	~1000	2000 - 6000

The sixth and last horizon comprises the unfrozen soil or rock beneath the permafrost with its characteristic resistivity (*f* in Figure 3).

In electrical measurements, the horizons 3 and 4 appear only as one layer with a very high resistivity.

This electrical cross section, as mentioned, is characteristic of a very thick, low-temperature permafrost. When the temperatures are high the cross section is very different (Figure 4). In this last case, depending on the nature of the rocks which constitutes the homogeneous geological media, on the minimal temperature and on the thickness of the permafrost, horizons 3 and 4 may disappear completely or be very thin (Figure 4).

When the bedrock is outcropping, the general shape of the cross section is maintained. The amplitudes, however, of the resistivity variations are much less than for Quaternary sediments. A relatively lower resistivity exists for frozen bedrock of any origin – sedimentary, metamorphic or igneous. The frozen bedrock compared to frozen sediments looks like a highly conductive media. Data obtained in the northeastern part of Siberia are shown in the table.

These data concern a region of low-temperature permafrost as thick as 60 to 200 meters. The ratio between frozen and unfrozen rock resistivities is about the same for schists and sandstones, approximately 12 to 20. For granites it is smaller – 2 to 6. These data are approximate, they are not universal; however they give a good idea of the ratio between frozen and unfrozen rock resistivities.

Such an electrical cross section depends obviously on the season. The first horizon does not exist in winter and the second horizon, where the resistivity varies abruptly, may also be modified by surface temperature conditions. The season, among other factors, need then be considered during resistivity interpretations.

Problems occurring during geophysical investigations of permafrost

There are two groups of problems which occur during geophysical investigations of permafrost.

1. The first one, not taking into account temperature conditions, is dealing with the usual geophysical difficulties that occur anywhere. However, one must remember that new boundaries separating media of different physical properties have been created in the homogeneous geological layer.

2. The second one concerns the different aspects of the physical study of permafrost in the geological environment which

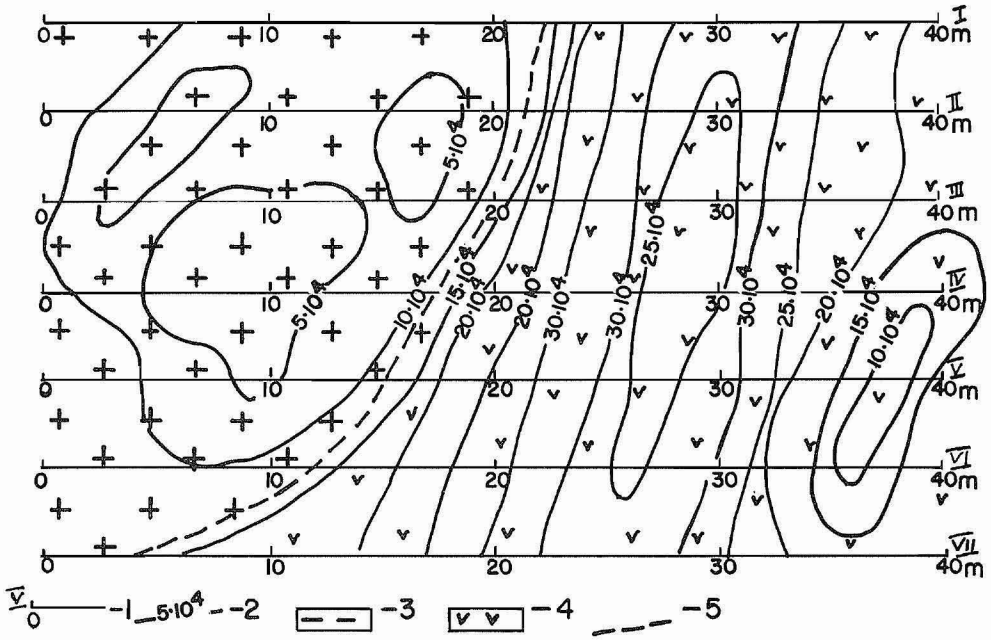


Figure 5. Resistivity map for an area covered with alluvial and eluvial deposits. 1 — electrical profiles, 2 — equiresistivity lines, 3 — area covered by eluvial deposits, 4 — area covered by alluvial deposits, 5 — boundary between alluvial and eluvial deposits.

may have a great influence in engineering design or water research.

When geophysical investigations are referred to only the electrical and thermal methods are considered because they are the ones that can be used on a large scale in permafrost regions.

Problems of the first group. Obviously, the problems to be solved by geophysical methods in permafrost regions are numerous and varied. According to our own experience, three types are worth citing here: (a) differentiation of Quaternary sediments and the measurement of their respective thickness, (b) geological mapping, stressing structural features, and (c) underground water studies.

The order in which these examples are described has been chosen according to their practical value and to the amount of data collected.

Differentiation of Quaternary sediments and the measurement of their respective thickness. A large amount of data collected allows us today to give a general picture of the resistivity of frozen sediments of different origins. These results come from electrical soundings above drill holes or mines and from direct measurements in holes and drifts.

Generally speaking, frozen alluvial deposits have high resistivities; some authors report from some thousands to 1 million ohmmeters.

The resistivity tends to increase when the percentage of small particles diminishes. In certain river valleys of the Chukotka Peninsula, for instance, it varies from 100,000 or 150,000 ohmmeters for sediments containing as much as 3% clay to 800,000 or 900,000 ohmmeters when this percentage decreases to one tenth of 1%. On high mountain plateaus, the stream deposits show higher resistivity than on hills.

In the valleys, the resistivity of the aluvium, frozen or not, varies more transversally than longitudinally. Stream deposits with equal resistivities stretch along the valleys; their width varies from 10 or 20 up to 100 meters and their length from hundreds to thousands of meters. The wider and older the valleys are, the wider is each equiresistivity area.

Among all the frozen Quaternary deposits, the glacial deposits show the highest resistivities: values up to 2 or 3 million ohmmeters have often been measured. Such results have been obtained, for example, for electrical soundings in the Chukotka Peninsula where there are numerous moraines which separate river basins. The contemporary frozen moraines in high mountains have lower resistivities and numerous specific geophysical properties. For instance, their resistivities vary from 10,000 to 40,000 ohmmeters. The eluvial and solifluction deposits show the lowest resistivities compared to all the other frozen Quaternary deposits.

Their resistivities are scarcely higher than 50,000 to 80,000 ohmmeters even in a low-temperature permafrost. These differences of resistivities between genetically different sediments allow their mapping. Not only have the absolute values of the measured resistivities to be examined but their distribution too. The resistivity map, Figure 5, is a good example.

The central part of the map shows very high resistivities; they could be either frozen alluvium or morainic deposits. The longitudinal pattern however, approximately along the axis of the old river, tells us that they are alluvium with varying grain size composition. The large decrease in resistivity in the northwestern section and the less marked drop in the eastern part show that these alluvia filling up the old valley are surrounded by eluvial clay deposits where the equiresistivity lines have lost their general longitudinal pattern.

One of the main problems for the geophysicist is to measure the thickness of these deposits. We know that the frozen bedrock shows a higher conductivity than the Quaternary frozen deposits: a diagram usually obtained for a vertical electrical sounding is of the K type ($\rho_1 < \rho_2 > \rho_3$).

In the northeastern part of Siberia, sometimes the flood plain deposits cover the stream deposits and the diagrams are then of

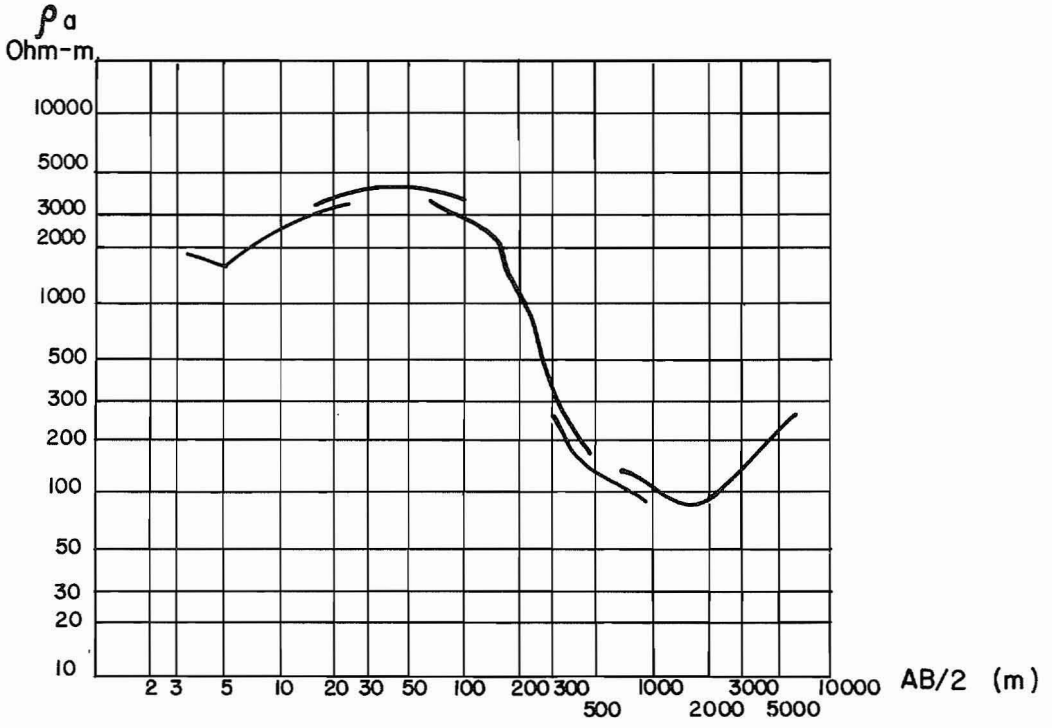


Figure 6. Electrical sounding diagram near Yakutsk.

the AK type ($\rho_1 < \rho_2 < \rho_3 > \rho_4$): this is characteristic of the Quarternary deposits in this area.

In areas where glacial and soft deposits of different age and great thickness can be found together, curves of the KQ type are obtained ($\rho_1 < \rho_2 > \rho_3 > \rho_4$). Certain types of electrical diagrams can be related to certain geomorphological features. A small-scale geomorphological map can then be used to predict the eventual result of electrical measurements and, on the other hand, electrical diagrams obtained in an area may be used by geomorphologists.

The measure of the thickness of the Quaternary sediments by electrical soundings is a complicated problem. The first difficulty arises because of the high ratio ρ_2/ρ_1. Secondly, the resistivity of the sediments, that is ρ_2, is not constant and shows large or small variations from one place to the other even for small dimensions of the array. At the same time the ratio h_2/h_1, the thickness of the frozen layer to the thickness of the active layer, varies as much. Besides these ratios ρ_2/ρ_1 and h_2/h_1 show seasonal variations because the thickness and the resistivity of the first unfrozen layer varies during the summer. Because of the variations of ρ_2/ρ_1 and h_2/h_1 in space as well as in time, one must often use the 'equivalence principle' when interpreting the sounding. KQQ ($\rho_1 < \rho_2 > \rho_3 > \rho_4 > \rho_5$) and KQ ($\rho_1 < \rho_2 > \rho_3 > \rho_4$) type curves in the case of low-temperature permafrost are hard to interpret.

In the field, to define the thickness of the Quaternary deposits, geophysicists usually compare the electrical diagrams to those that have been obtained above drill holes or wells. An empirical correlation is established between the position of the horizons and the points on the curve where a variation occurs. On uniform and homogeneous sediments, this procedure leads to satisfactory results.

Consideration has been given to the idea of studying the electrical diagrams by transforming the cross sections artificially (Yakoupov, 1959). A small variation of ρ_a^{max} allows us to give to the media responsible for the maximum, the chosen resistivity of permafrost. This way only a few resistivity values have to be used for the intermediate layer.

When numerous soundings have been made in a given area, statistics are first used to determine the values of the maxima of the apparent resistivity. This way, the values of ρ_2 used in the interpretations are known. The precise values are obtained on very thick layers of soft sediments. When the thicknesses are less than 10 m the errors can be great.

Geological mapping of the bedrock and resolution of structural problems. The fact that there are supplementary boundaries inside the permafrost creates considerable difficulty for the interpretation and the structural mapping. After 1930 when the geophysicists were first dealing with permafrost, they considered this high-resistivity superficial layer as an impassable obstacle. Now it has been proven that, in some cases, the geology beneath the permafrost can be studied.

In Central Yakutia, for instance, as reported by Kaleunov (Kaleunov, 1958) deep electrical soundings allowed for the study of very important and isolated structural features. In this area, the resistivity of the frozen rocks varies from 100 to 10,000 reaching sometimes 25,000 ohmmeters or more. The thickness of the permafrost varies from some tens to some hundreds of meters. On the diagrams this layer appears as a sharp maximum of ρ_a (Figure 6). Then the resistivity decreases from a layer of Mesozoic argillaceous sands beneath the frozen layer. The final portion of the curve ascends with a 45° slope which marks the top of the Oldhamian massif or the Archean basement. In this

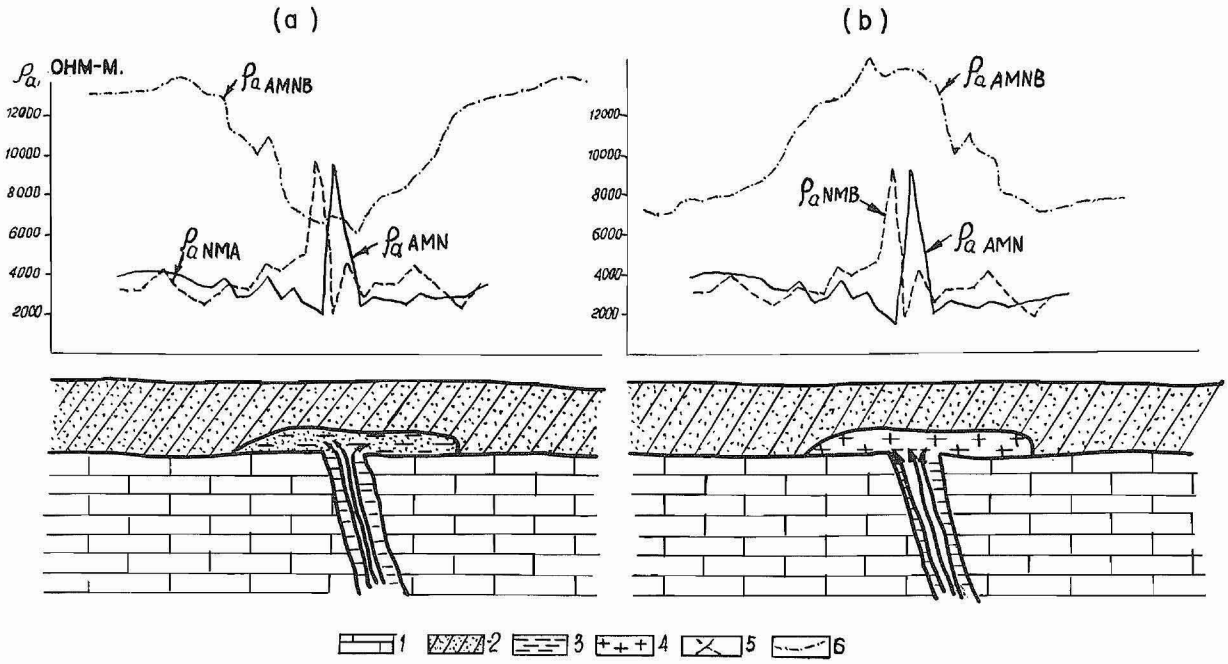


Figure 7. Electrical profiles over water channels, (a) for thawed soft deposits, (b) for fossil ice in the soft deposits. 1 – bedrock, 2 – soft Quaternary deposits, 3 – thawed area, 4 – glacier, 5 – electrical profile with asymmetrical arrays with 3 electrodes *AMN* and *NMA* with a large distance *AM*, 6 – electrical profile with symmetrical arrays *AMNB* with small *AB*.

case, the structural feature appears around 800 to 1000 m and the frozen layer has a very small influence on the position of the last ascending branch. However, when the top of the structural horizon is close to the bottom of the frozen layer, the electrical methods are not very successful. The use of 'electrical profiling' for structural or general geologic mapping, for locating specific layers or tectonic features, encounters many difficulties (Akimov, 1959). The distance between the current electrodes is very hard to choose. This distance has obviously to be very large to get information on the layers underneath the permafrost. The fact that these dimensions have to be increased diminishes very much the resolving power of the electrical profiling compared to an investigation in unfrozen areas. Because of that and because of the inhomogeneities of the frozen layer, thin layers cannot be distinguished. Tectonic features are hard to map beneath the permafrost as well as inside it. Beneath the permafrost, the same difficulties as for the thin layers arise. Inside the permafrost, these features may be noticed only if they create special channels for the water through this frozen layer.

Underground water studies. Our experience in using geophysics to study hydrogeological problems in permafrost is very small. That is why we will speak only of the few problems for which we have some data.

The channels of the water that enter the permafrost from underneath can easily be detected. The host rock surrounding these channels becomes unfrozen which produces a large drop in resistivity. The amount of unfrozen rock and hence the size of the ρ_a anomaly is related to the depth and to the water flow. If the water is flowing through cracks in rocks the width of the unfrozen zone is related to the thermal conductivity of the rocks. In soft permeable sediments, the arrival of this heat thaws the ice in the pores. The presence of such a zone around the water channel creates a large area of low resistivities. However, in the bedrocks these anomalies have smaller widths than in soft sediments. The fact that the ratio between the resistivities of frozen and unfrozen material is much higher for the sediments than for bedrocks is also important.

On the resistivity diagrams, the water channel in the bedrock and the unfrozen area surrounding them appears as an anomaly of the 'conductive sheet' type with very high conductivity, small thickness and small extension. Such anomalies require the use of an asymmetrical array (3 electrodes of dipole array) with a very small spacing between the measuring electrodes. In the soft sediments, these unfrozen spots give birth to wide anomalies that can be located with a symmetrical as well as an asymmetrical array (Figure 7a). Two different arrays with two different depths of penetration can actually be used, one locating the anomalies in the bedrock and the other studying the electric behaviour at the bottom of the Quaternary sediments.

Often the water circulating in the bedrock discharges into the sediments and gives birth to fossil ice. These can be located as very high maxima on the resistivity diagrams (Figure 7b). Under certain conditions, a reciprocal behaviour may be noticed between the two arrays, one with large spreads, and the other with smaller spreads. A minimum for the large array due to the water channel may be related to a maximum on the small array due to a fossil ice in the sediments.

On the electric well logs, the anomalies are often huge, even if the drill holes do not intersect the water channels but go only through the surrounding thawed area.

An SP anomaly coincides with the ρ_a log because of the electrofiltration phenomena and of the potential differences created at the boundary between the two phases of the water.

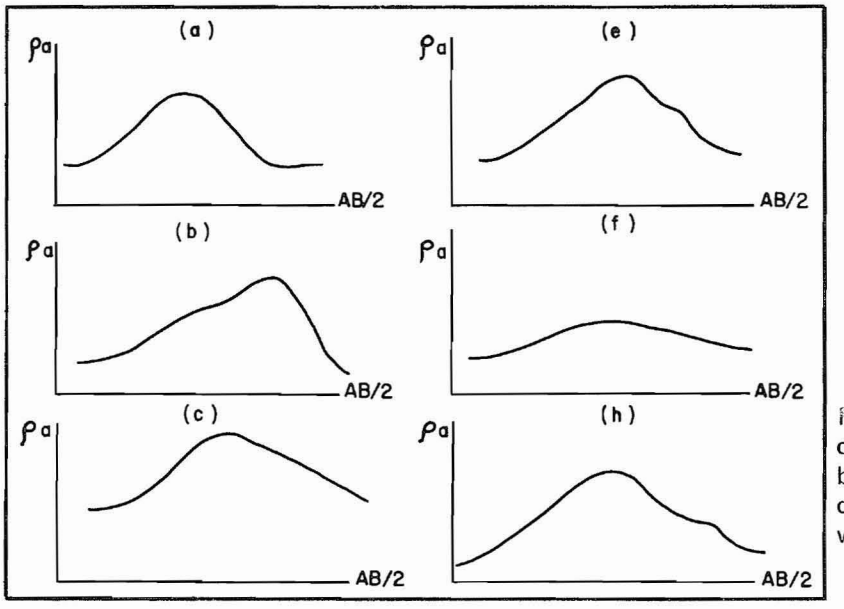


Figure 8. Typical electrical sounding diagrams in permafrost. (a,b,c) The lower boundary occurs within the Quaternary deposits. (e,f,h) This boundary occurs within the bedrock.

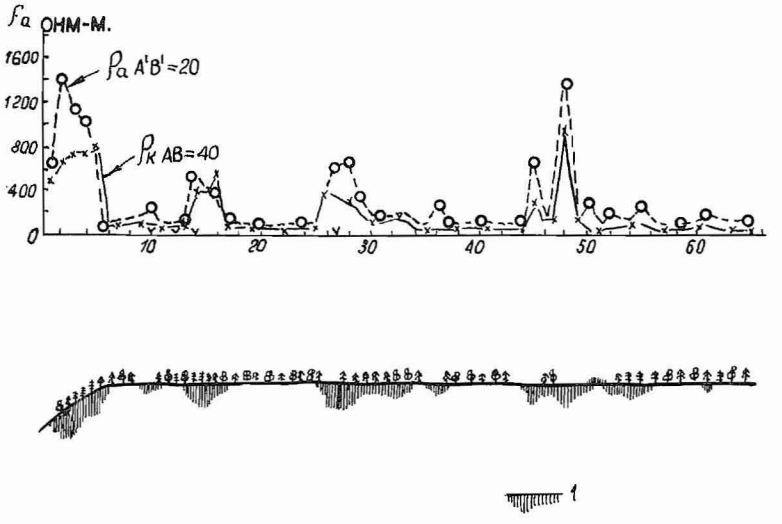


Figure 9. Electrical profiles with AA^1 MNB^1B arrays (from Dostovalow, 1959). 1 — frozen soils.

One can predict that new and numerous geoelectric and geothermic investigations will be applied to the hydrogeology in permafrost regions.

Problems of the second group. As said previously the second problems are directed towards the study of the permafrost phenomena itself. First, the specialists are interested in the boundaries of the frozen soils, the lower and upper surfaces, its thickness, the presence of unfrozen areas or warm water in the media, the mapping of the extension of the permafrost, of fossil ice masses and the changes of these surfaces with time.

Most of the time, electrical soundings are used to study the boundaries of the permafrost. The principle of the method has been largely explained in the above section where the electric cross section of the permafrost and the measure of the thickness of the sediments were described.

The upper boundary of the permafrost can always be defined. It is shown by a rapid increase of the resistivity in the diagram.

The errors in this case, are due to the gradual increase of the resistivities from the unfrozen to the frozen soils. For pure sands, the boundary is sharp. When clay and sand deposits appear together, the limit is not as sharp because they freeze at different temperatures.

The definition of the lower boundary is more complicated. If the soft sediments are very thick and if the lower boundary of the permafrost occurs within them, it appears as the top of the bedrock would. When permafrost is very thick, however, the Quaternary sediments are rarely homogeneous. Some typical diagrams are given in Figure 8. The electrical inhomogeneities often lead to great difficulties in interpretation when establishing these boundaries. For instance, when alluvial and morainic deposits with their high resistivities cover frozen eluvial deposits which are more conductive it is hard to define the lower boundary when it occurs within the latter. In such cases, the lower boundary occurs as a change in the slope of the right descending branch of the curve (type KQ). When the boundary occurs inside the bedrock and when it is covered by frozen sediments, its definition is as difficult as in the previous case especially when the sediments are of alluvial or morainic origin.

To interpret these soundings and define this lower boundary, the artificial transformation of the cross section is used as in the case of the studies of the Quaternary sediments.

In other cases, the electrical soundings permit the discovery of thick, thawed areas within the permafrost.

From the definition of the lower and upper boundaries of the permafrost a map of its extensions can be drawn. For instance, the analysis of many soundings lead E.N. Kalenov to the conclusion that in Yakutsk the permafrost is more highly developed on the right flank of the Lena River than on the left. Soundings made on the ice on the river show the lack of permafrost underneath; the large volume of flowing water contains so much heat that it cannot develop.

However, an electrical sounding among the electrical prospecting methods is a relatively expensive procedure and it is not absolutely necessary for the mapping of the extent of permafrost. Often with considerably less expense the electrical profiling method can be successfully used. Preferably such measurements have to be carried out together with visual checking of permafrost (Dostovalow, 1959). The profiles are chosen so that they cross the main terrain features which are related to the permafrost structure. In this way, with a less expensive technique, more data is obtained. On the profiles, the frozen rocks appear as maxima (Figure 9). When the mapping of the extent of the permafrost is carried out in regions of low-temperature soft-frozen sediments the profiles are easily interpreted. The difference between the resistivities of unfrozen and frozen sediments (-5° or -8°C) is such that the errors due to possible changes in lithology are negligible.

The main difficulties arise from studies in high-temperature regions of isolated frozen areas where the difference in resistivity between frozen and unfrozen rocks is about the same as the difference due to lithology. In such problems in the case of clay or Quaternary deposits many observations have to be taken. At the same time, special interpretation techniques have to be used in conjunction with the drill-hole control.

The study of permafrost includes the tracing of the shift of its boundaries because of natural conditions and man's activities.

To follow such variations, periodic measurements have to be taken. The resistivity maps in the vicinity of inhabited areas show the gradual shift in the high-resistivity zones in relation to the recession of the frozen rock boundary. The measurement of resistivity in drill holes permits the tracing of the thawing of the rocks around underground tunnels. The efficiency of frozen screens used in some engineering projects can be controlled. In all these cases, the temperature and resistivity measurements can be taken on the surface of the earth as well as in the drill holes.

Conclusion

Geophysical work in permafrost regions is carried out on a broad scale. The DC resistivity method has been used almost exclusively up to the present. The use of the AC current method looks promising because the high resistivities of frozen rocks do not constitute an obstacle as they do for the DC current method.

Reference

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