TERRESTRIAL HEAT FLOW STUDIES AND INVESTIGATIONS OF GEOTHERMAL RESOURCES IN INDIA

A THESIS

Submitted in fulfilment of the
requirements for the award of the degree
of

DOCTOR OF PHILOSOPHY

in

GEOPHYSICS

By

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CANDIDATE'S DECLARATION

I hereby certify that the work which is being presented in the thesis entitled "Terrestrial Heat Flow Studies and Invêstigations of Geothermal Resources in India" in fulfilment of the requirement for the award of the Degree of Doctor of Philosophy, submitted in the Department of Earth Sciences of the University is an authentic record of my work carried out during a period from 1972 to July, 1980 under the superbision of Prof. V.K. Gaur and Dr. Hari Narain.

The matter embodied in this thesis has not been submitted by me for the award of any other degree.

Mohan L. Gupta
Candidate's Signature

DE

This is to certify that the above statement made by the candidate is correct to the best of our knowledge.

VK Gaur

Date: Signatures of Supervisors

CERTIFICATE

The thesis presented by Shri M.L. Gupta embodies the results of investigations carried out by him since 1972 under our supervision and guidance. We certify that this work has not been submitted for the award of any other degree or diploma.

Hari Narain | V.K. Gaur
(Hari Narain) | (V.K. Gaur)

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CHAPTER V GEOTHERMAL RESOURCES OF INDIA

ABSTRACT

As is well known, the earth is a great reservoir of thermal energy which continuously flows outward towards its surface at the rate of about 3.1×10^{20} calories per year, which is considerably larger than the total energy annually released through volcanoes and earthquakes. Occassionally, this is manifested at the surface in the form of hot springs geysers, fumaroles and volcanic eruptions.

Geothermal studies shed significant light on the evolution of the earth and the nature of heat and mass transfer in its interior and various other deep seated processes. In recent years they have attracted considerable attention owing to their diagnostic value in delineating plate kinematics and hot spots; in particular, regions of anomalously high geoheat which could be exploited as a replenishable resource in an increasingly power hungry world.

Investigations aimed at studying the thermal state of the crust in the Indian subcontinent through investigations of terrestrial heat flow at a large number of carefully selected locations, and exploration of the geothermal resources of some of its hot spring areas forms the subject matter of this thesis. Results of these investigations and their analysis, synthesized with

other geodata especially generated for this study as well as that already available are discussed.

TERRESTRIAL HEAT FLOW STUDIES

Heat flow determination involves accurate measurement of subsurface temperature gradients at various depths in boreholes and a detailed study of the thermal conductivity of concerned rocks. Equipment used for these measurements were specially designed and fabricated for this purpose. These are briefly described in the thesis.

Heat flow values including thermal logging data obtained in a large number of boreholes at various locations in the following tectonic units of the Indian subcontinent are discussed, as well as values of the coefficients of thermal conductivity of rocks.

1. The Cambay basin
2. The Koyna region
3. The Aravalli Mountain Belt
4. The Dharwar Schist belt

Terrestrial heat flow values thus obtained in various parts of the Indian subcontinent show considerable variations which may be summarised as follows:

* The Dharwars and the areas covered with the rocks of Aravalli and Pre-Aravalli Super Groups are characterised with low heat flow values ($\approx$ 0.95 HFU)

as are the Deccan Traps which most probably overlie the eroded shield rocks of low radioactivity, the Upper Mantle temperatures under low heat flow areas of the Indian Shield being most likely around 400°C.

* Proterozoic miogeosyncline belts and areas of platform deposits are distinguished in their thermal character by significantly higher heat flow values of 1.7 HFU.

* The part of the Cambay basin located towards the north of the Mahi river also shows high heat flow values of 1.8 - 2.3 HFU, the main cause for which appears to be the transient thermal perturbations produced by an extensive injection of basaltic material in the crust during the Upper Miocene-Pliocene times.

Besides, an analysis has also been attempted of the various Precambrian Shields of the world which are generally found to be associated with low heat flow values. Some anomalously high heat flow values have been observed in the Indian (Aravalli, Cuddapah basin etc.) as well as in other shield areas. These variations apparently related to the evolutionary history of shields and shield segments, when examined for their global occurrences, appear to suggest that the Laurasian Shields are generally associated with lower heat flow values than those of the Gondwanaland.

* 1 HFU = 10^{-6} cal.cm^{-2}.sec^{-1} = 41.87 mW m^{-2}.

The results of heat flow investigations have been further discussed in respect ofsome peninsular regions, in the light of their known geology, tectonics and near surface concentration of heat generating radioactive elements, such as was available. There appear to be lateral variations in the sub-surface temperatures within the lithosphere underlying the Indian Shield and these may cause thermal stresses responsible for some of the seismic activity along ~~xx~~ ancient zones of weakness.

EXPLORATION OF GEOTECHNICAL RESOURCES

India is a subcontinent with a large population, where local conditions and availability of natural resources vary greatly from one region to another, particularly of energy resources which constitute a crucial input for economic and social development. The only way to provide energy to every corner of the country, whether in the inhospitable desert regions or in the remotest regions of perennial snows, without wastage and without degrading the environment, is to develop and exploit repleneshible sources of energy. Geothermal resources belong to this unconventional group. Attempts to explore and develop geoheat sources both of the natural hydrothermal systems in hot spring regions and also of dry hot rocks, have been initiated in a number of countries and intensified in others where some exploration was already in progress.

In order to gain a better insight into the geothermal resources of India and to recognise, if possible, certain geothermal provinces where subsurface thermal waters may be encountered and exploited, an analysis and synthesis has been attempted of available geodata particularly of the surface manifestations of geoheat and available heat flow data in the Indian Peninsula in the light of the major tectonic features. It is felt that non-volcanic geothermal resources of moderate and low grades can be expected to occur in certain geothermal provinces of India.

Pursuing encouraging results of the reconnaissance work for geothermal exploration in the Puga valley (Ladakh), detailed investigations were planned and carried out to establish the character and extent of the geothermal sources of the valley. The results of geothermal investigations so obtained i.e., estimation of the natural heat loss from the surface, probable reservoir temperature, sub-surface thermal conditions, structure of the Puga valley and its power potential, as well as their analysis is contained in a later chapter of this thesis. Additionally, results of some investigations in the Parbati valley (Manikaran-Kasol) hot spring area have also been included.

While summarising the main results of the present study, some problems of basic and applied nature have also been delineated which need to be investigated in further details. Suggestions have also been included for more detailed exploration in the promising areas.

CHAPTER I

INTRODUCTION

1.1 Geothermal Studies - A Preview

Surface manifestations of geoheat in the form of hot springs and geysers and the occurrence of volcanoes, long associated in men's minds with supernatural phenomenon, furnish us with direct evidence that the earth is a veritable reservoir of thermal energy. We now know that heat flows continuously from the earth's interior towards its surface at the rate of approximately 1×10^{13} calories per sec. Most of this is lost through conduction, whilst a small part is dissipated by volcanoes, earthquakes and other tectonic processes.

The transfer of the earth's internal heat towards its surface, constitutes a veritable thermodynamic engine which is responsible for various geodynamic and associated physico-chemical phenomena e.g., the earth's magnetism, and its rhelogical and geochemical structure. Furthermore, the distribution in space and time of magmatism, diastrophism, lava flows and of ore concentrating solutions essentially reflect the prevailing characteristic patterns of heat generation and of heat and material transport in the earth during the corresponding epoch. The most dramatic expressions of these processes are, however, exhibited in the earth's crustal features. For, the brittle spherical shell of the lithosphere constrained to glide over a softer asthenosphere tends to translate the effect of stresses produced by underlying thermal convection cells into thermally and kinematically compatible lithospheric plate movements which are, in turn, responsible for a host of geodynamic phenomena such as earthquakes, Island Arcs and mountain building, crustal thickening, folding, faulting and various emplacements of differentiated materials.

A study of the extant thermal state of the upper part of the earth's crust which is accessible to us, against the backdrop of other geological, geophysical and geochemical information of the region, can therefore be quite illuminating in working out the progression of various earth processes in time, leading upto the present context with all its implications towards the elucidation and generalization of global processes. Besides, geothermal studies can be extremely valuable in delineating geothermal resources which could be tapped to produce a perennial and environmentally clean source of energy.

The present work is concerned with investigations of the thermal state of the Indian crust through measurements of ~~x~~ subsurface temperatures and terrestrial heatflow made at a large number of widely distributed locations. It also includes geothermal exploration of some hot spring areas of the Indian land mass and synthesis of various geodata towards evaluating their resource potential.

1.2 Sources Of Geoheat

The surface heat flow depends on the underlying sources of heat in the interior, their magnitude and distribution. During the course of the evolution of the earth over the last 4-5 billion years, the direct means of internal heat generation have varied. The principal contribution (over 80%) at present is assumed to arise from the decay of long lived isotopes of Uranium, Thorium and Potassium which are distributed in varying quantities in the earth's interior and its crust. Other sources of heat in the past, could well have been furnished by i) the gravitational energy i.e., energy released during the formation of the earth's core, ii) tidal friction.

The earth is believed to have been formed out of a primordial gas cloud which trapped a considerable amount of gravitational energy during its accretion. Part of this, later appeared as heat during the consolidation of the primitive planet towards its present layered structure enclosing a high density core.

The actual mechanism of accretion of particles and the radiative balance obtaining at that time is not well understood, which makes it difficult to estimate the gravitational contribution to the earth's heat except that such effects might have been considerable. Lubimova (1967) estimated a figure of 1.5×10^{38} ergs, arising from this source.

Another significant source of heat is furnished by tidal friction. Munk and McDonald have estimated that nearly 3.2×10^{19} erg/sec of the mechanical power of the earth-moon system is dissipated in the earth. The relative proportions of this energy dissipated in the sea and in the solid earth are not settled, but McDonald finds enough evidence to assign a major part of this to the solid earth. If, as it is believed, the moon was nearer to the earth in earlier times, the tidal energy dissipated in the earth must indeed have been quite considerable.

Another major source of energy is furnished by chemical transformations particularly exothermic, which occur inside the earth leading to magmatic and volcanic activities. Extensive geological field studies should be of help in estimating the actual extent of such activities in space and time.

Radioactive heat generation is contributed both by shortlived as well as by longlived isotopes, the former being largely active in the initial stages of geological history. From a consideration of the present heat flow of the earth and heat generation in chondrites, Birch (1960) and others have indicated that radioactive distribution in the earth might be similar to those in chondrites. On this assumption, workers have assumed average concentrations of Uranium, Thorium and Potassium as being 1.1×10^{-8} g/g, 4.4×10^{-8} g/g; and 8.0×10^{-8} g/g respectively. The decay rates for U^{238}, U^{235}, Th^{232}, and K^{40} are respectively 1.54×10^{-10} yr^{-1};

$9.71 \times 10^{-10}\ yr^{-1}$; $0.499 \times 10^{-10}\ yr^{-1}$ and $5.5 \times 10^{-10}\ yr^{-1}$. Lubimova's(1967) calculations show that heat production inside the earth may reach a high value ranging from 0.6 to 2.0×10^{38} erg; i.e., an average of 1.4×10^{38} erg depending on details of the earth's composition during its entire life (4.5×10^{9} years).

1.3 Geothermal Energy Resources

Although heat is generally available in all subsurface rocks as manifested by the average temperature gradient of about 30°C per km, it is mostly too diffuse to be exploited economically. However, in certain regions anomalously large amount of heat is stored in the upper subsurface layers. Such conditions are generally encountered i) in young orogenic belts where igneous activity has occurred during the last few million years, ii) in zones of crustal rifting and iii) near margins of lithospheric plates and regions which have experienced extensive crustal movements in recent times. In the latter circumstances, magma is often generated and injected upto depths of several kilometres from the surface, thereby creating a heat source at that level, upon solidification. Such regions are characterised by large thermal gradients and are termed as 'Geothermal Areas', as they are potential regions for harnessing the thermal energy. Meteoric waters percolating into the earth through permeable fractures and fissures upon reaching such heated rock regimes, in turn, get heated and may form a reservoir of steam, hot water or a mixture of both.

The reservoirs are usually self revealed by the presence of thermal springs or/and geyser activity, and could be tapped through boreholes drilled to appropriate depths, to provide hot water for a variety of industrial and cultural uses.

For practical purposes, geothermal reserves can be defined as the total quantity of heat which can be economically extracted and used. The present day technology, makes it economically feasible to extract geoheat for electricity generation from and upto depths of about 4 km provided that the thermal energy so tapped is considerably greater than average, as happens to be the case in regions of anomalously high heat flow.

Geothermal areas are generally found close to the plate boundaries (Figures I-1) which are also associated with earthquakes and occas~~s~~ionally with volcanic activity.

1.3.1 Exploitation of geothermal resources

Man's exploitation of geothermal resources dates back to prehistoric times, when he used natural hot springs for bathing, cooking and for therapeutic purpose. Gradually, many of the well known geothermal areas of the world were turned into recreation centres equipped with baths and swimming pools. In countries like India they became places for pilgrimage and worship. Belief in the miraculous power of the natural thermal waters to cure diseases and to improve health generally persists even to this day, and has also

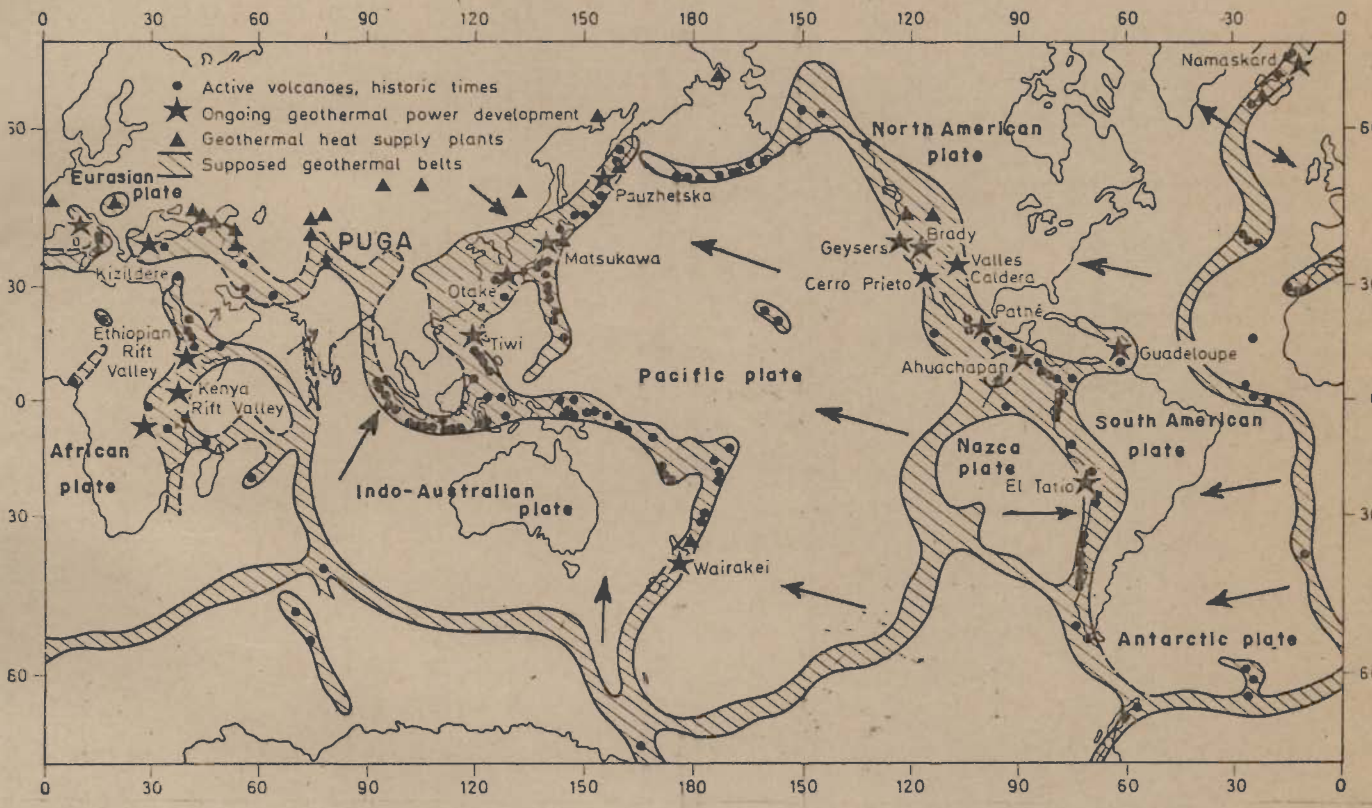


FIG. I-1 Plate Boundaries and Possible Geothermal Regions of the World attracted medical attention for applications in balneotherapy, particularly in Japan and the USSR. In colder climates, naturally occurring hot water was used to warm houses and public buildings even centuries ago. The English town of Bath was built by the Romans around thermal springs two millenia ago. Iceland has used geothermal heat since the 13th century when Snorri Sturlson, a saga writer heated the walls of his sod and lava rock house with the waters from a thermal spring. In arid regions, hot spring water and condensed steam has been used for dinking and other domestic purposes. A variety of materials like sulphur, boric acid, Kaolinitic clays, mercury and alum have been recovered from fumaroles and hot springs for centuries.

Systematic exploitation of geothermal resources on a large scale was first started in Larderallo in Italy in 1812 with the recovery of boric acid from the fumaroles. The first experimental generation of electricity using a natural steam also started at Larderallo in 1904 when a local count, Picrro Ginori Conti upon some misunderstanding with the local power company, decided to experiment with natural steam as an energy source for power generation. Over the next ten years the technology was sufficiently well developed to enable generation of 250 kilowatts of geothermal power commercially.

The Italian experiment promoted other countries also to explore their geothermal resources and today geothermally powered electricity generating plants operate in a number of other countries, notably, New Zealand, Iceland, the Soviet Union, Japan, Mexico, USA, El Salvador and Turkey. Additionally, a number of countries make non-electrical use of geothermal energy (Figure I-1) for space heating and in process industries, and world wide interest in its development is rapidly gathering force. The present status of geothermal energy utilization for power globally, is shown in Table I-1. Table I-2 outlines the minimum temperatures needed for various applications of geothermal energy.

Exploitation of geothermal energy is relatively free from the dangers of environmental pollution. It does not require complex technologies as nuclear power installations do, nor massive construction works as in the case of hydroelectric systems. Techno-economic studies of geothermal power system, made in various countries indicate that geothermal power is comparatively much cheaper, costing only about a third of that obtained from fossil fuels or nuclear sources.

1.3.2 Investigations of geothermal resources in India

Accounts of thermal springs and their therapeutic uses are recorded in the Puranic and Buddhist literature of ancient India. But scientific interest and work on Indian hot springs began in early 19th century, Ludlow (1826); Wilson (1827); Turner (1828); Prinsep (1831, 34); Wade (1837) and Duncan (1838). Thomas Oldham (1882) computed the first inventory of the location and temperature of 301 hot springs of

TABLE I-1

Global picture of existing and programmed Geothermal Power (1975)

Name of country	Name of the Geothermal Field	Installed generating capacity (Megawatts)	Plants under construction (Megawatts)
United States	The Geysers (Dry steam)	502	406
Italy	Larderello (Dry steam)	380.6	
	Travale	15	
	Monte Amiata	22	
New Zealand	Wairakei	190	120
	Kawerau	10	-
	Broad lands	-	125
Japan	Matsukawa (Dry steam)	22	68
	Otake	13	
	Onuma	10	
	Onikobe (Dry steam)	25	
	Hatchobaru	-	50
	Takinoue	-	50
Mexico	Cerro Prieto	75	75
El Salvador	Ahuachapan	-	90
Iceland	Namafjall	2.5	-
	Krafla	-	60
Philippines	Tiwi	-	100
	Laguna	-	110
Soviet Union	Pauzhetsk	5	-
	Paratunka	0.7	-
Turkey	Kizilders	0.5	11.5
	Total	1,273.3	1,265.5

TABLE I-2

Required temperature of geothermal fluids for different applications (after Lindal, 1973)

°C		
180	-	Evaporation of highly concentrated solutions. Refrigeration by ammonia absorption digestion in paper pulp, kraft.
170	-	Heavy water via. hydrogen sulphide process, drying of diatomaceous earth.
160	-	Drying of fish meal; Drying of timber.
150	-	Alumina via. Bayer's process.
140	-	Drying farm products at high rates. Canning of food.
130	-	Evaporation in sugar refining. Extraction of salts by evaporation and crystallisation.
120	-	Fresh water by distillation. Most multiple effect evaporations, concentration of saline solution.
110	-	Drying and curing of light aggregate cement slab.
100	-	Drying of organic materials, seaweeds, grass, vegetables etc., washing and drying of wool.
90	-	Drying of stock fish. Intense de-icing operations.
80	-	Space heating. Greenhouse by space heating.
70	-	~~Animal husbandry.~~ ~~Farm~~ Refrigeration (lower temperature limit)
60	-	Animal husbandry. Greenhouses by combined space and hot bed heating.
50	-	Mushroom growing. Balneological baths.
40	-	Soil warming
30	-	Swimming pools, biodegradation, fermentations, Warm water for year round mining in cold climates, de-icing.
20	-	Hatching of fish. Fish farming.

India and adjoining territories. Since then information on Indian hot springs has been gradually accumulating. Ghosh (1948), Warring (1965) and Chatterjee (1969), Hot Spring Committee (1968), over 250 centres of hot springs are presently known to occur in the Indian subcontinent. Important sites are shown in Figure I-2.

Scientific interest towards the development of hot springs areas of India, as possible sources of energy came to the fore during the last two decades. The 'Hot Spring Committee' set up by Government of India and the National Committee on Science & Technology, initiated a systematic conceptualization and implementation of promising areas of geothermal research. This activity received an added stimulus in the wake of the oil crisis which burst on the scene in 1973. As a result, several programmes for the identification, assessment and development of geothermal resources were taken up, as well as exploration of geothermal resources in significant hot spring areas, those marked for the first intensive investigations being:

1. Puga, Ladakh.
2. Manikaran, Parbati valley, H.P.
3. Konkan coast, and
4. Sohna, Haryana

Besides, a reconnaissance of the hot springs of the Beas valley, NW Himalayas, Monghyr-Rajgir, Assam; Godavari valley, Rajasthan etc., has also been made by various authors (vide bibliography of relevant publications and reports, Annexure - I).

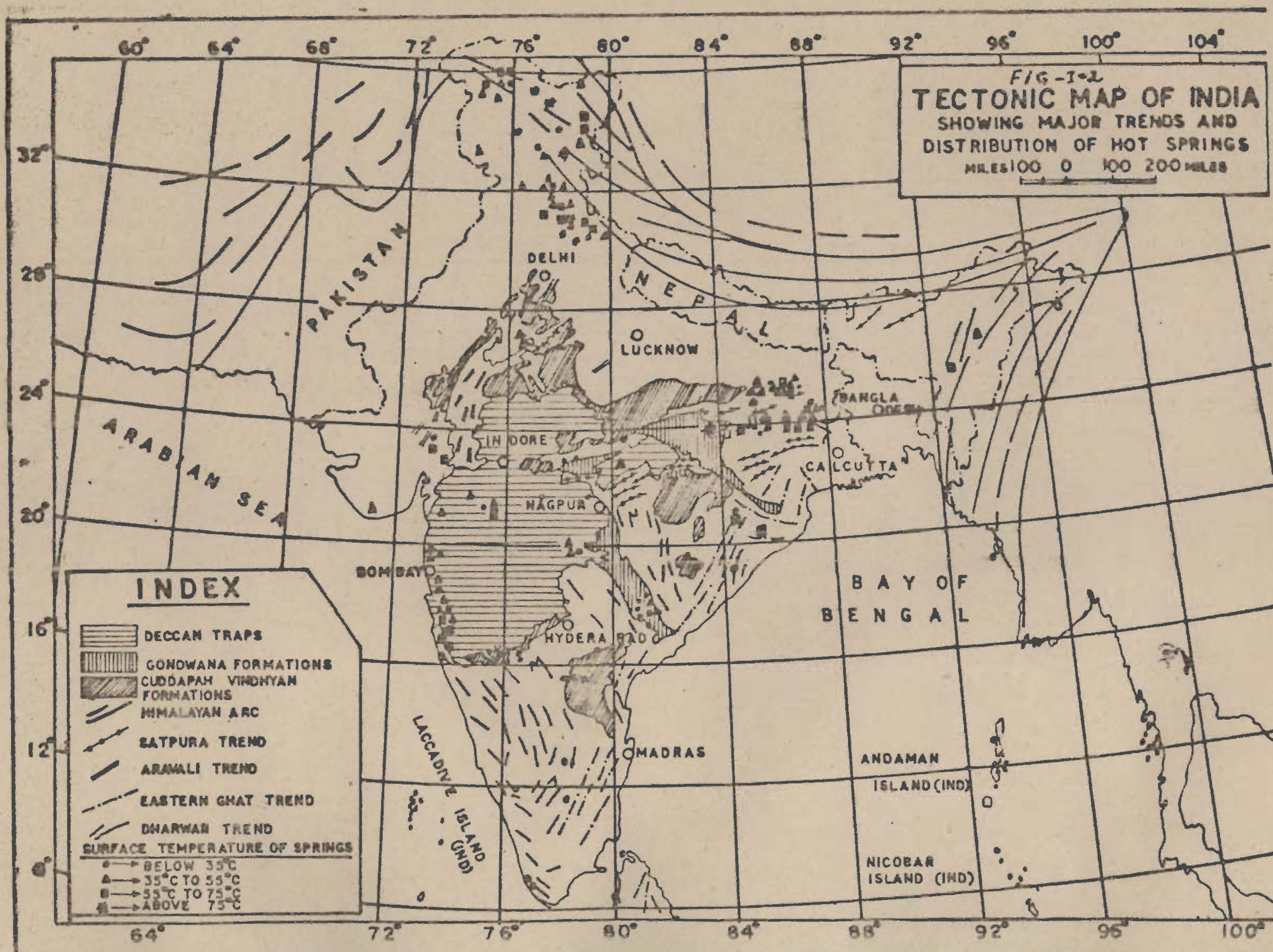


FIG-I-2
TECTONIC MAP OF INDIA
SHOWING MAJOR TRENDS AND
DISTRIBUTION OF HOT SPRINGS
MILES 100 0 100 200 MILES
INDEX
DECCAN TRAPS
GONDWANA FORMATIONS
CUDDAPAH VINDHYAN FORMATIONS
HIMALAYAN ARC
SATPURA TREND
ARAVALI TREND
EASTERN GHAT TREND
DHARWAR TREND
SURFACE TEMPERATURE OF SPRINGS
BELOW 35°C
35°C TO 55°C
55°C TO 75°C
ABOVE 75°C
PAKISTAN
ARABIAN SEA
DELHI
NEPAL
LUCKNOW
BANGLA DESH
INDORE
CALCUTTA
NAGPUR
BOMBAY
HYDERABAD
MADRAS
BAY OF BENGAL
LACCADIVE ISLAND (IND)
ANDAMAN ISLAND (IND)
NICOBAR ISLAND (IND)
60° 64° 68° 72° 76° 80° 84° 88° 92° 96° 100° 104°
32° 28° 24° 20° 16° 12° 8°

1.3.3 Types of geothermal resources

Seven distinctive types of geothermal resources can be recognised. These are:

1. Dry steam reservoirs - Such reservoirs contain superheated steam at a high pressure trapped in confined aquifers, which can be tapped by drilling and fed to turbines directly. Known dry steam reservoir systems are found in Geysers (USA), in Larderello (Italy), in Matsukawa and Onikobe (Japan). Such reservoirs are almost non-existent in India, but Puga geothermal field may produce dry steam on development.

2. High temperature water reservoirs (150 - 250°C). Such reservoirs produce a mixture of vapour and hot water under pressure. In India, such reservoirs are found mostly in the Himalaya, although those of Agnigundala (Godavari valley) and Tatapani in Surguja district, M.P., may probably also fall in this category. In such reservoirs steam is seperated and even the fluids are flashed to a lower pressure thereby turning them into steam suitable for powering turbines. However, this involves considerable wastage of heat.

3. Low temperature water reservoir (60 - 150°C). Such reservoirs are the most common, being found in almost

all geological units. Most of the hydrothermal systems of Peninsular India are of this type. Their thermal waters are most suitable for use in industry and for space heating, agricultural and horticultural production and for power generation through secondary turbines.

4. Heat in normal areas. Subsurface temperature gradients from about 10°C to 30°C/km occur in normal areas which cover a major part of the continents. In principle, this heat can be extracted for use but would necessitate very deep drilling to reach ambient high temperatures. Technology to accomplish this is still not economically feasible but may prove to be a viable source in the future.

5. Geopressurised resources. These are pressurised water reservoirs within sedimentary basins capable of supplying both heat and mechanical energy together with dissolved methane. Examples are, Gulf Coast of the United States and the Cambay basin (India) reservoirs.

6. Hot dry rock. Cooking intrusive bodies at about 600°C but without a heat exchanger (water) are found to occur in various places over the continents although the absence of hydrothermal manifestation make it difficult to locate them. This type of resource is quite promising and can be developed by hydrofracturing of hot crystalline rocks and induced circulation

of water through the fissures so created. Attempts in this direction are being made by the Los-Almos laboratory of USA.

7. Magma resources. Large quantities of heat may also be tapped from shallow magmatic regions. R&D programmes towards accomplishing this are being seriously considered in Japan and USA.

1.4 Terrestrial Heat Flow Studies

Heat flow determinations being a direct indicator of the earth's thermal state, have steadily increased since the first reliable measurements made in England (Benfield, 1939), and in South Africa (Bullard, 1939), to over 7,250 values now available over the continents and oceans (vide table I-3).

TABLE I-3

Growth of Global Heat Flow Determinations

Year	Number of Heat Flow data set over the		Total	Reference
	Continent	Oceans		
1954	43	20	63	Birch (1954)
1963	73	561	634	Lee (1963)
1965	131	913	1,044	Lee & Uyeda(1965)
1972	255	2,329	2,584	Sclater (1972)
1976	1,699	3,718	5,417	Jessop et.al (1976)
1980	2,800	4,400	7 7,200	

A significant discovery which emerged from these measurements quite early was that the average heat flow over continents and ocean floors were almost equal (Bullard et.al., 1956). This was contrary to expectations for it was believed that the oceanic crust being deficient in heat producing radioactive elements should exhibit a lower heat flow value. However this unexpected results was explained by Sclater (1972). Later values of high average heat flow over oceans results from upward biassing of heat flow values over the continental average as most of these came from the anomalously high heat flow regions in the oceans i.e., the oceanic ridges overlying spreading centres.

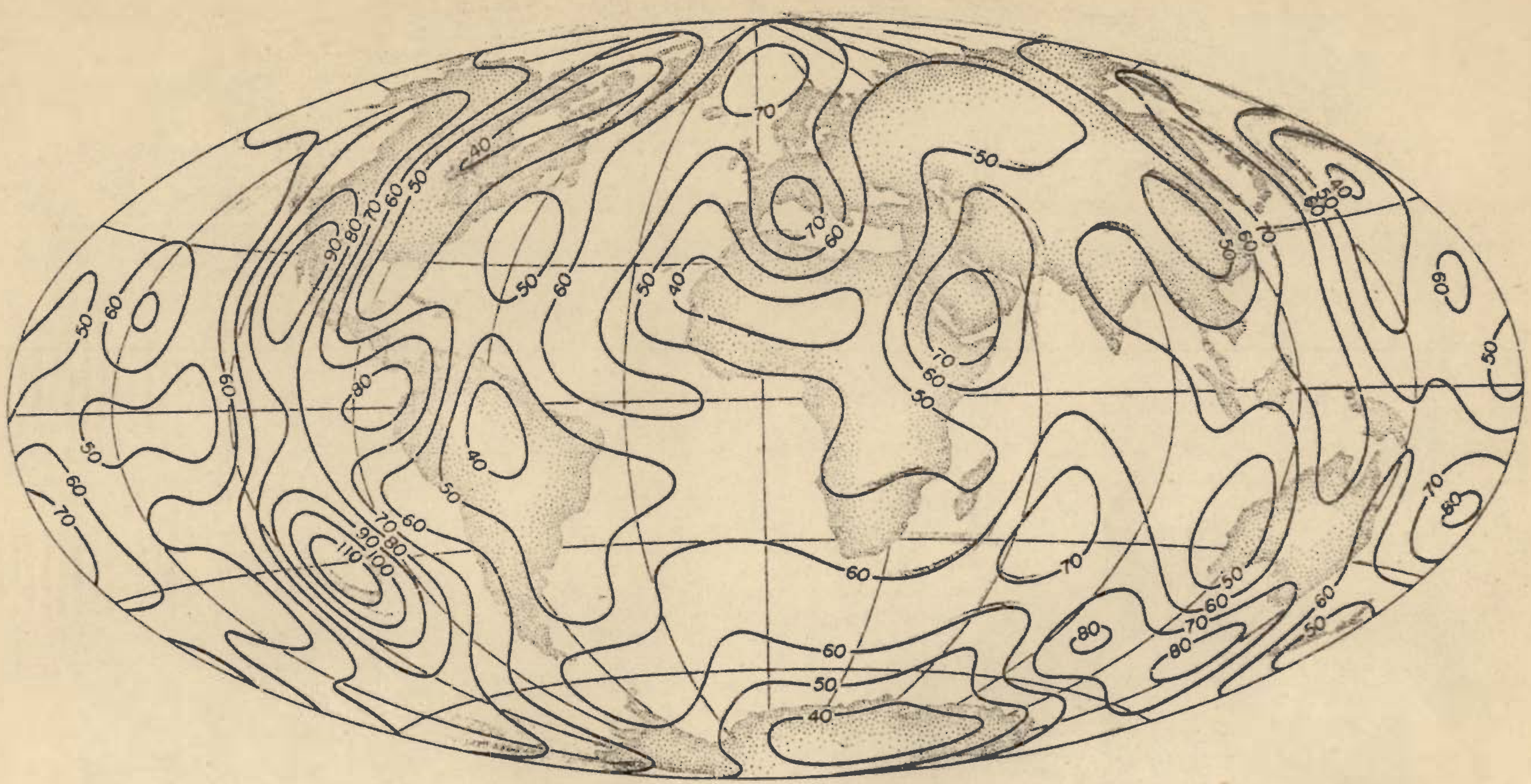
With steady increase in the number of heat flow values over different sections of the globe, several correlations clearly emerged. Thus,continent heat flow was shown to be correlated with the age of a geological or tectonic province,being lower for older provinces, Birch (1954), Kraskovskiy (1961), Lee and Uyeda (1965), Gupta (1967) and Polyak and Smirnov (1965), Hamza and Verma (1969) found similar results in respect of the age of basement rocks and heat flow values measured over them. Subsequently it was recognised that continental heat flow values in a region was inversely proportional to the last tectono-thermal event suffered by it. The correlation of heat flow with age of the ocean floor or of the basaltic crust magmatically generated at and transported away from the oceanic ridges was also later established by Sclater and Francheteau (1970) and Mckenzie and Sclater (1971). It is interesting to note that the average heat flow value over both continental and oceanic regions finally decay to approximately the same mean value i.e., around 1.1 HFU, observed in the ocean basins and over Precambrian shields, although the decay processes and the decay constants for the two cases happen to be different. The variations of heat flow values observed over continents are mainly the result of initial differentiation of the underlying crust and mantle, erosion and cooling of perturbations introduced by the last tectono-thermal event whilst in the case of oceanic regions, the decrease of heat flow with age is mainly controlled by the residual heat of cooling of the oceanic lithosphere according to $Q = at^{-\frac{1}{2}}$ upto approximately 120 m B.P. whereafter an equilibrium value is reached.

Effect of erosion and uplift on the redistribution of heat producing elements and on subsurface temperatures and heat flow, have been discussed by various workers. Tien-Chang Lee (1980) demonstrated that the transient heat flow introduced by erosion reaches the equilibrium value after an elapse-time (equal to 20% of the total main duration of erosion. Vitorello and Pollack (1980) using heat flow data over continental regions of different age groups as given by Chapman and Furlong (1977), advocated a three component model for the decrease of continental heat flow with tectonic age. Their first component of approximately 40% of the observed heat flow in terrains of all ages originating from radioactive disintigration in the top crustal layers. This top layer being exposed to erosional processes, depletes radioactive isotopes from it due to erosion, appears to follow a ponential time scale resulting in decrease of heat flow with age. Various workers have supported the view that major erosion, after the initial development of continental land mass, occurs during the first 300-400 m.y. resulting in removal of the radioactive isotopes from the crustal isotopic enriched zone. Victorello and Pollack suggest that the second component nearly 30% of the observed heat flow arises from the residual heat associated with tectonic activization of the area. This component gradually decrease to zero for late to mid Precambrian terrains. The last component of observed heat flow comprising 30% constitutes a back ground heat flux originating within the conductive root zone coupled with the heat energy flowing outwards perhaps from the core. A magnitude of 27 mW/m^2 (0.64 HFU) was suggested for the third component which should be unvarying beneath all continental terrains. Exceptions to these generalizations however occur over Archaean segments of the planet Earth which are characterised with heat flow values of 27 mW/m^2 or even lower. Another point still open to speculation is the time scale of the outward flow of the deep heat.

The linear relationship between continental heat flow and heat production in near surface rocks as given by Birch et al. (1968) has been discussed in a later chapter in the thesis. This relationship helped to distinguish and delineat heat flow provinces, each of which is associated with characteristic heat flow Q_o (reduced heat flow) originating beneath the top enriched radioactive layer. Values of Q_o decrease as one moves from younger tectonic provinces to old Archaean provinces. For example Q_o is found to be about 69 mW/m^2 for the Basin and Range province (Lachenbruch and Sass, 1978) and decrease to as low as 22 mW/m^2 under the Baltic Shield, 23 mW/m^2 under Dharwar Craton India (this thesis chapter IV B), 11 ± 8 mW/m² under Niger North Western African Craton (Chapman and Pollack, 1979). Some discussion of the reduced heat flow and mantle component of heat flow is given in chapter IV C of the thesis.

Global heat flow maps based on the spherical harmonic analysis of available and interpolated global heat flow data have been produced by various workers from time to time, Lee and MacDonald (1963), Lee and Uyeda (1965) Horai and Simmons (1969), Lubimova and Sujetnova (1975), Chapman and Pollack (1975).

One such map given by the latter is reproduced here in Figure I-3. A spherical harmonic analysis up to the 12th degree yield a global mean value of 59 mW/m^2. A recently revised map (Chapman and Pollack, 1979) by using about 1800 additional heat flow values gave a mean heat flow value over the continents (including marine shelves) and oceans as 60 and 95 mW/m^2 respectively and a global mean equal to 81 mW/m^2.



Degree 12 spherical harmonic representation of global heat flow using observations supplemented by predicted heat flow where no observations exist. Heat flow in mW m^{-2}.

1.4.1 Continental heat flow studies in India

Continental heat flow studies at the National Geophysical Research Institute commenced during 1963. The work started with measurement of temperature in short narrow horizontal tunnels in mines. With the fabrication of vertical borehole logging equipment during 1965, temperature measurements in exploration boreholes were also commenced. Thermistor probes for temperature measurements and part of the equipment for the determination of coefficient of thermal conductivity of rock samples were designed by the author.

Review of Indian heat flow studies summarising upto date research efforts and our current state of knowledge have been given by Verma et al. (1968), Gupta and Rao (1970), Verma and Gupta (1975) and Gupta et al. (1979). Information regarding the thermal field of various geotectonic segments of the Indian land mass have been discussed in a later chapter in this thesis.

1.5 Motivation and Scope of Present Work

Heat flow investigations carried out in the Khetri Copper belt (Aravallis) and in Cambay gas field of the Cambay Basin, during 1967-68, showed high heat flow values of 1.76 (HFU) and 2.3 (HFU) respectively, contrary to the expectation that the Pre-Cambrian shields were characterised by low heat flow values. These findings triggered considerable interest and raised quite a few questions.

According to Krishnan (1953) the Aravallis after continuing under the Cambay basin branch into two, one branch extending into the Arabian Sea and the other continuing under the Deccan Traps and merging with the Dharwar trend. Naturally question arose as to whether the heat flow anomaly in the Cambay basin was contributed both by the Aravallis and the Deccan Traps or wholly by perturbations caused by Deccan Trap volcanism which may have tapened into the Miocene-Pliocene? What were the magnitudes of surface heat flow in other parts of the Cambay Basin where different thicknesses of sediments overlay the Deccan Traps and differential block movements occurred during its evolution? Were the observed axial gravity high and the high heat flow of the Cambay basin related in some way? Did they originate from a Common source, that is from a basic magmatic intrusion? In order to resolve some of these questions it was felt that a good number of heat flow values not only in the Aravallis and Cambay basin, but also in some places in Deccan Traps and even in the area lying towards south of Deccan Traps must be obtained. Thus a programme of carrying out heat flow studies from the Aravallis to the Dharwars through the Cambay Basin and the Deccan Traps was concepturized. Equipment used for accurate measurements of subsurface temperatures at various depths in drill holes and for the determination of the coefficient of thermal conductivity of rocks was adapted, redesigned and fabricated, which is but briefly described in Chapter II. Subsequent heat flow measurements in other parts of the Indian Shield Verma et al., 1969; Gupta et al. 1970; Rao and Rao, 1974; Gupta, 1980; Chapter IV A, this thesis, showed large variations. In order to better understand such variations, an analysis of the heat flow data of the Precambrian Shields of the world has also been attempted. This work along with thermal logging data obtained in a large number of boreholes and values of coefficients of thermal conductivity of different rock types encountered, as well as heat flow values are discussed in chapters III, IV A, IV B and IV C, together with other available geodata.

In order to gain a better insight into the geothermal resources of India and to recognise, if possible, certain geothermal provinces where subsurface thermal waters may be encountered and exploited. An attempt has also been made to re-examine and analyse the available heat flow data and information pertaining to the evolution of different tectonic units of the Indian land mass including the occurrence of hot springs and character of thermal waters (Chapter V).

Pursuing encouraging results of the reconnaissance work for geothermal exploration in the Puga valley (Ladakh), detailed investigations were planned and carried out to establish the character and extent of the geothermal resources of the valley. The results investigations so obtained, as well as their analysis is contained in Chapter VI of this thesis. Additionally, results of some investigations in the Parbati valley (Manikaran-Kasol) hot spring area including the inferences drawn from a combined interpretation of the data gathered by various workers for the Parbati valley geothermal field have also been included. While summarising the main results of the present study and reviewing the present status of heat flow studies in the three cratons viz., Aravalli, Dharwar and Singhbhum ~~xx~~ of Indian land mass some problems of geothermal importance both of basic and applied importance have also been delineated for further detailed investigations. Some suggestions have also been made for more geothermal exploration in these areas (Chapter VII).

CHAPTER II

MEASUREMENT TECHNIQUES, EVALUATION METHODS AND SURVEY METHODOLOGIES USED IN THE PRESENT WORK

2.1 Measurement Of Terrestrial Heat Flow

For the determination of terrestrial heat flow at a location, two quantities need to be measured viz., the geothermal gradient and the coefficient of thermal conductivity of the rock formation in which the temperature gradient has been measured. Temperature measurements within the earth can be traced back to a reference by Morin (1619). But the first systematic measurements of land heat flow values were made in South Africa and Great Britain by Bullard and Benefield during 1939. However, measurement techniques and the instrumentation used for field and laboratory measurements as well as evaluation methodologies have undergone considerable development and change since then, of which a review is given by Beck (1965). The methods and apparatus used, for the present studies, for determining the coefficient of thermal conductivity of rocks and for the measurement of underground temperatures are described below.

2.1.1 Measurements of temperature and temperature gradients

Temperatures were measured in boreholes at discrete depth intervals using calibrated thermistor thermometers connected through a three-conductor Vector Cable to a Wheatstone bridge to compensate for the resistance of the lead cables. The probe-cable assembly passes through a portable and manually operated pulley system to which a revolution counter was attached (Figure II-1). Three types of thermistor probe assemblies were used. One of these consisted of a set of 12 thermistors housed and sealed in three stainless steel tubes of length about 15 cm and diameter 2.5 cm, each tube containing four thermistors. All the 12 thermistors were so connected as to have a total resistance of about 4,000 ohms at 0°C and 1,400 ohms at 25°C. The assembly was constructed and calibrated by Dr. F. Roy at Harward University, Cambridge, Massachusetts. This probe dissipated approximately 10^{-4} watts without self heating through more than 0.001°C. It has been successfully used in areas where temperatures did not exceed 60°C. Additionally, two other probes each using a single Fenwal GB34P91 thermistor having a resistance of about 4,000 ohms at 25°C were specially fabricated and calibrated for these studies. One of these (Figure II-2) was housed in a thin diameter stainless steel tube whilst the other was enclosed in a brass tube. They had a wider range of temperature but dissipated lesser heat. They were usually calibrated and checked both before and after each field trip. Although, absolute temperatures were not directly used in the determination of the heat flux, temperatures measured by these probes are accurate to within 0.02°C. The temperature gradients were usually accurate to within a few percent for small intervals of about 10 m and to less than 1% for intervals of the order of 100 metres.

FIG. II-1 Equipment for measurement of Geothermal Gradients, showing thermistor probes, cable, Pully system and the bridge.

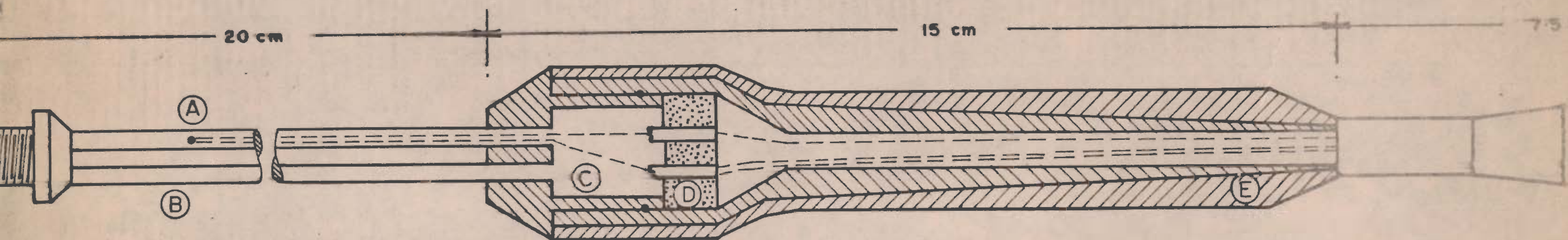


FIG. I-2 Thermistor Probe Details

2.1.2 Measurement of thermal conductivity

Various methods have been reportedly used for measuring the coefficient of thermal conductivity of rocks and building materials. In respect of the present studies the divided bar method was largely used, for which Benfield's (1939) apparatus was modified to improve its precision and performance. The conductivity of brass bars was determined by calibrating it against crystalline quartz. However, several shortcomings were encountered in using the original Benfield's apparatus. Measurements had to be made on three or more discs to eliminate the effect of contact resistance, involving a lot of time, both in the preparation of discs as well as in making actual measurements. Further, errors also crept in due to a slight variation in the values of conductivities of the various discs even when made from the same core sample due to various factors viz., inhomogeneous mineral composition, nonuniform contact resistance and 'parallel' arrangement of minerals with decreasing disc thicknesses. The guard-ring divided-bar apparatus described by Beck (1957), though better in performance and accuracy, suffers from the same inherent shortcomings as those of the Benfield's apparatus. Another type of divided-bar apparatus which was described by Birch (1950), involves six measurements on a single disc, the contact resistance being eliminated by the application of large axial pressure through a hydraulic jack. In order to reduce the time required for a measurement and improve its precision and performance, several modifications were introduced in Birch's apparatus during the present studies. The electric heater was replaced by a water jacket which sprayed water from a thermostatically controlled bath over the top end of the stack of discs that sandwitch the sample. A similar water jacket was provided at the lower end of the stack of discs. The water jackets were insulated from a pair of hemispherical thrust bearings by 6 mm thick perspex discs. Thermostatically controlled water at 40°C and 35°C or at 35°C and 30°C was circulated through the jackets.

With the above mentioned modification, the apparatus becomes a constant temperature difference type of divided-bar apparatus, and the time to reach equilibrium is greatly reduced. The water jackets were coated with a film of silcon grease to minimise contact resistance, and to affect a uniform temperature distribution over the copper discs containing the differential thermocouples. In order to reduce lateral heat losses, U-foam or cotton insulation was used around the specimen and the reference discs. The lateral heat losses from the stack of discs were further minimized by enclosing the apparatus containing the thrust-bearings and the stack assembly in a steel shield whose temperature was maintained at the mean temperature of the two water jackets. Finally, to prevent rapid exchange of heat from the steel shield with the environment, the whole apparatus (Figure II-3) was encased in a wooden-glass enclosure.

For purposes of a calibration, a sample of Lexan, which is a polycarbonate plastic, supplied by Prof. Birch of

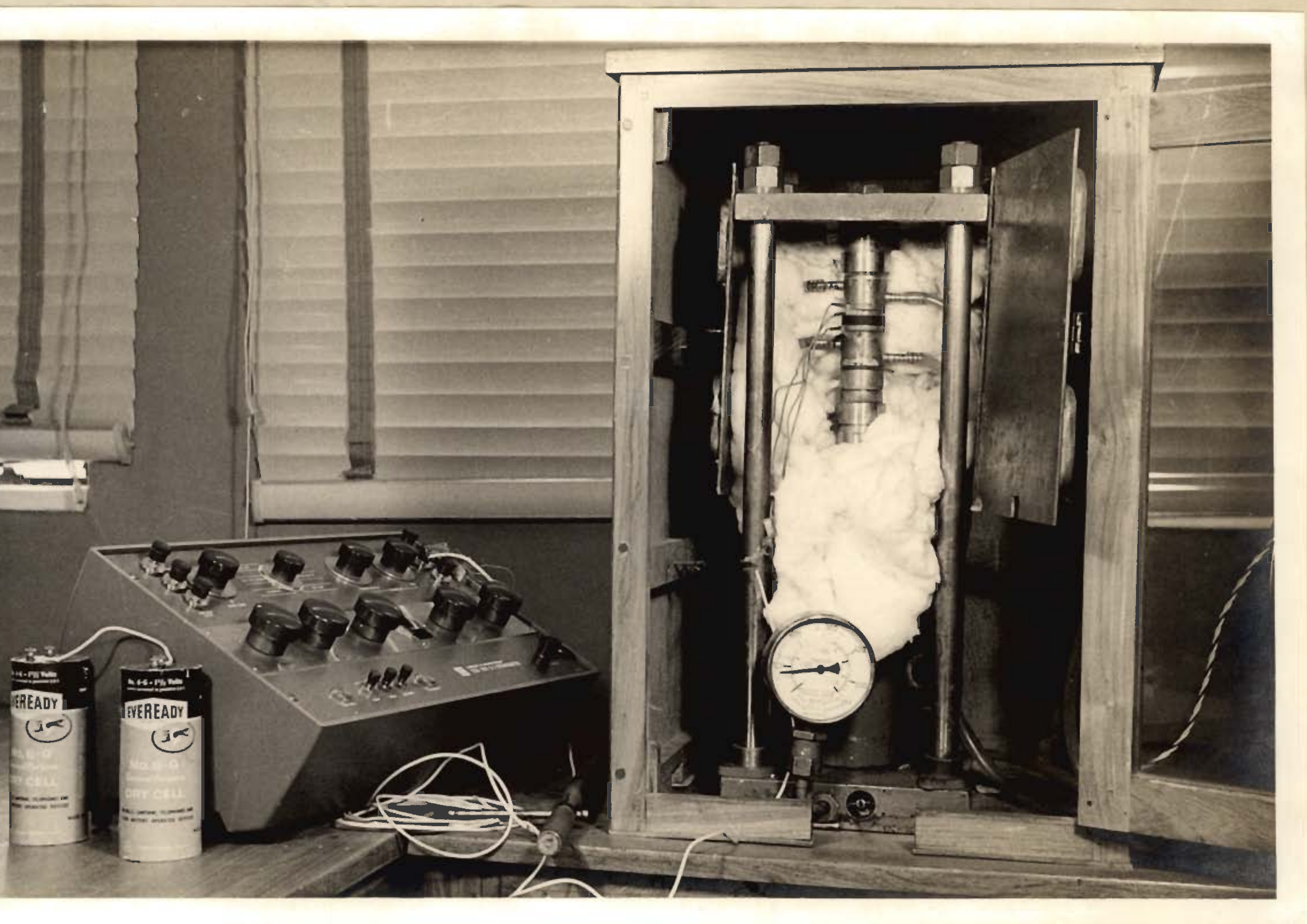


FIG. II-3 Thermal Conductivity Apparatus

Harvard University, was used as a reference material, by first calibrating it against quartz. This was necessiated owing to frequent damage of quartz samples when used directly as a primary reference.

The specimens were prepared from cut or cored samples in the form of cylindrical discs of diameter 4.12 cm or 2.54 cm and thickness of about 1 cm. The flat faces were ground parallel to within 0.03 mm and flat to within 0.01 mm.

The specimen discs were saturated with water or oil under vacuum and coated with films of silicon grease before assembling the stack. The stack consisted of: i) a sample disc; ii) two discs of a reference material of known conductivity, and iii) four copper discs all of equal diameters in the following order : Copper disc → disc of reference material → → copper disc → sample disc → copper disc → disc of reference material and copper disc. Copper and constanton thermocouples were carefully inserted in the copper discs for measurements of temperature gradients across the reference and specimen discs. On assembling the stack into a proper position it was subjected to an axial pressure of about 100 kg/cm^2. After the system reached stability, temperature differences were measured using copper - constanton differential thermocouples connected to a K-type potentiometer and a galvanometer of sensitivity of 0.46 μV/mm deflection at one metre distances. This modified apparatus takes about 12-15 minutes for attaining steady state conditions and was used for practically all measurements of the coefficient of thermal conductivity reported here.

Conductivity coefficients of sediments collected from shallow depths in various hot spring areas were measured on well preserved samples sealted in water tight containers by a transient radial heat flow method, using an axial linear heat source. The probe consists of a hypodermic needle of 20 gauge with an advance heating wire and a copper-constanton thermocouple or thermistor enclosed in it (Figure II-4). The needle is heated by the internal heater at a known constant rate, and the rise in its temperature measured by a sensitive recorder. The needle, which is of a small diameter compared with its length (length to diameter ratio is about 70 : 1), can be regarded as a line source of heat which is fully energized within a few seconds of the heater power being turned on. Details of the system and the experimental procedure have been described elsewhere (Gupta, 1960).

2.2 Calculation of The Coefficient of Thermal Conductivity

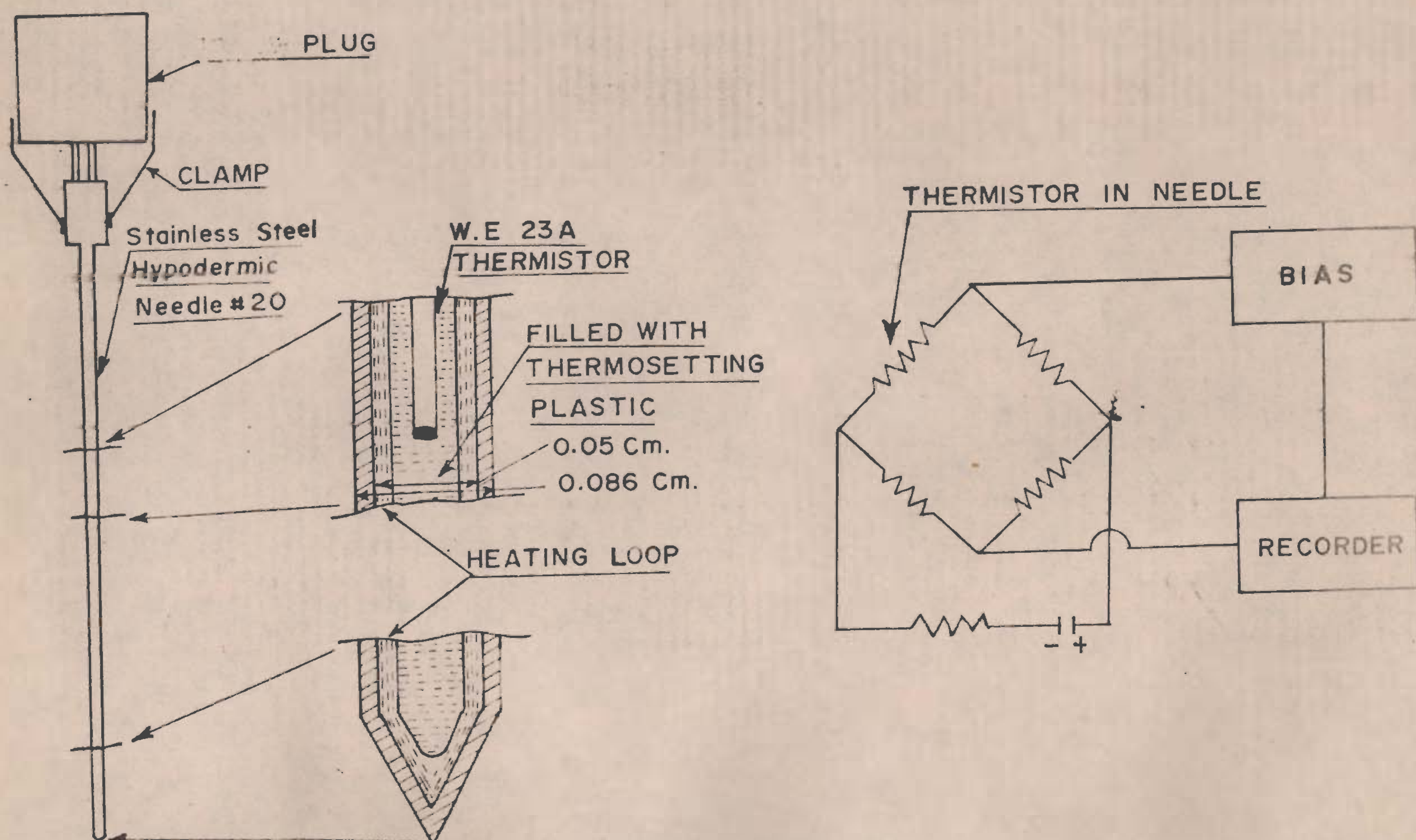
2.2.1 Steady state method

If an equal quantity of heat is assumed to flow through the entire cross sections of the stack consisting of the reference and sample discs, the coefficient of thermal conductivity K_s of the specimen can be expressed as follows:

$$K_s \frac{\Delta T}{D_s} = K_r \frac{(\Delta T_1 + \Delta T_2)/2}{D_r} \qquad \ldots \text{ II-1}$$

FIG. 4

NEEDLE PROBE FOR CONDUCTIVITY MEASUREMENTS



where K_r is the conductivity coefficient of the reference material. D_s and D_r are the thicknesses of the specimen and reference discs and ΔT, ΔT_1 and ΔT_2 are the respective temperature gradients across the specimen and the two reference discs.

$$\text{or} \qquad \frac{K_s / D_s}{K_r / K_r} = \frac{\Delta T_1 + \Delta T_2}{2 \Delta T} \qquad \ldots \text{ II-2}$$

2.2.2 Transient radial heat flow method

The basic mathematical treatment of transient heat flow in materials has been given by Ingersoll, Zobel and
1959 Ingersoll (1948), and Carslaw and Jaeger (1948).

The Fourier equation for heat conduction due to an infinite line source having azimuthal symmetry can be written as:

$$\frac{d\theta}{dt} = \alpha \left[\frac{d^2\theta}{dr^2} + \frac{1}{r} \cdot \frac{d\theta}{dr} \right] \qquad \ldots \text{ II-3}$$

The above equations when solved for a special case, where heat is flowing radially at a constant rate from an infinitely long line source embedded in an infinetely extending medium, initially at zero temperature, yields 'θ' the temperature at any point in the infinite medium; through the following equation:

$$\theta = \frac{Q}{2\pi K} \left[I \left(\frac{r}{2\sqrt{\alpha t}} \right) \right] \qquad \ldots \text{ II-4}$$

where,

K = Coefficient of thermal conductivity of material surrounding the line source.

Q = heat input per unit time from a unit length of the source.

r = distance from heat source.

t = duration of heating reckoned from the beginning of experiment.

and the series $I\left(\frac{r}{2\sqrt{\alpha t}}\right)$ is given by

$$I(x) = C - \log_e x + \frac{x^2}{2} - \frac{x^4}{8} + \ldots\ldots$$

$$x = \frac{r}{2\sqrt{\alpha t}}$$

when $\frac{r}{2\sqrt{\alpha t}}$ is negligibly small,

$$I(x) = C - \log_e x$$

and

$$\theta = \frac{Q}{2\pi K}\left(C - \log_e \frac{r}{2\sqrt{\alpha t}}\right) \qquad \ldots \text{ II-5}$$

when all temperatures are measured at the line source i.e., 'r' is fixed

$$\theta = \frac{Q}{4\pi K}(\log_e t + C') \qquad \ldots \text{ II-6}$$

Also, C and C' are constants. The temperature rise between times t_1 and t_2 is given by

$$\Delta\theta = \frac{Q}{2\pi K}(\log_e(x_1) - \log_e(x_2))$$

$$= \frac{Q}{2\pi K} (\log_e \sqrt{\frac{t_2}{t_1}})$$

$$= \frac{Q}{4\pi K} (\log_e \frac{t_2}{t_1}) \qquad \ldots \text{ II-7}$$

According to equation II-6 the plot of θ against $\log_e t$ therefore results in a straight line whose slope is = $Q/4\pi K$, whence 'K' can be calculated if 'Q' is known.

Differentiating equation II-6 with respect to time we get

$$\frac{d\theta}{dt} = \frac{Q}{4\pi K} \frac{1}{t} \qquad \ldots \text{ II-8}$$

The reciprocal of $d\theta/dt$ when plotted against 't' gives a straight line passing through the origin. In actual practice this straight line cuts the X-axis at a point $t = t_o$. This is ascribed to the initial time taken for the heat to flow into the surrounding conducting medium on account of the finite diameter of the probe. The slope however can again be utilized to yield the value K.

2.3 Evaluation Of Heat Flow

Computations of the vertical component of the heat flux Q_z are based on the integration of the fundamental equation of steady heat flow along the z direction.

$$Q_z = - K \frac{\delta T}{\delta z} \qquad \ldots \text{ II-9}$$

where K and ~~$\delta T/\delta z$~~ $\partial T/\partial z$ are respectively the conductivity coefficient and the temperature gradient, and Z is the depth measured vertically downwards.

2.3.1 The 'GK' method

In the simplest case Q_z is obtained by the product of the least square gradient and the mean conductivity coefficient obtained from the measurements of a suitable number of samples from the corresponding depth section and is denoted by 'GK'. This can be used for areas consisting of massive rocks which do not exhibit any systematic variation of the conductivity coefficient or temperature gradient with depth. When more than one long linear sections are encountered on the temperature profile, heat flow is calculated separately for each section and finally a mean value is obtained.

2.3.2 The 'BP' method

In layered rocks, long linear sections are seldom obtained on the temperature profile as the conductivity coefficients exhibit a wide variation. A good value of Q_z in such cases is obtained by integrating equation (II-9), Bullard (1939), to give

$$T_z = T_o + Q_z \int_0^z \frac{\Delta Z}{K} \qquad \ldots \text{ II-10}$$

where T_o is a constant.

As temperatures and conductivities are measured in discrete intervals, the above integral can be replaced by the sum $\Sigma_i \Delta Z_i . R_i$, where ΔZ_i is the section of the borehole penetrated by the rock of conductivity coefficient $K_i = 1/R_i$.

Equation (II-10) gives a linear relation between the measured quantities T_z and $\Sigma_i \Delta Z_i . R_i$. Q_z is obtained from a least-square determination of the slope of these quantities. Any change in heat flow with depth is indicated by the departure from linearity of the above plot. Residuals from the least square line through all the measured points, are then examined for any systematic variations. In actual practice, values of $\Sigma_i \Delta Z_i . R_i$ are obtained from the lithologic information, together with conductivity measurements on representative core samples. This method has been often used, and is denoted by 'BP'. A simpler method, which is a combination of the above two methods is also used occassionally. First, the least-square temperature gradients are calculated for those depth intervals within which the curves are substantially linear. Each formation is then weighed according to its conductivity value and thickness obtained from stratigraphic information. Finally, the weighted mean conductivity values of specific depth intervals are computed, which when multiplied by the temperature gradient yields the desired heat flux.

2.3.3 The 'MBP' method

The above methods are strictly valid only for horizontally layered homogeneous structures and yield the best mean value for the flux flowing vertically through the rock immediately surrounding the borehole. In case of beds dipping appreciably from the horizontal, the lines of flow will suffer refraction at the boundaries resulting in preferential heat through rocks of higher conductivity. No complete theory for such a case has yet been developed. In practice, observations are reduced on the tacit assumption that structures encountered are largely horizontal, Jaeger (1965). Roy (1963) by assuming geometrical refraction at the boundaries of parallel dipping beds, suggested approximate relations for obtaining the true heat flux that would be expected if the structures were homogeneous. Equation (II-10) when modified (Hyndman and Sass, 1966) to take the geometrical refraction into account under the assumption that the heat flux vector eventually becomes vertical at some depth where layered homogeneous materials are encountered, becomes:

$$T_z = T_o + Q_o \left[\frac{\cos^2\theta_o}{K_i} + \frac{\sin^2\theta_o}{K_o} \right] \Delta D_i \quad \ldots \text{ II-11}$$

where D_i and K_i are the thicknesses and conductivities of the dipping beds penetrated by a vertical borehole, θ_o is the dip angle of the beds, and Q_o is the flux in the substrate which are horizontal and possess a conductivity coefficient equal to K_o.

In order to evaluate heat flow by this method a reasonable value for K_o has to be assumed. This becomes more or less possible in areas of known geology. The method has been used in a few cases, and is denoted by 'MBP'.

2.4 Equipment For Shallow Thermal Exploration And For Studies For Natural Heat Losses

Shallow subsurface temperatures at depths of about one metre have been measured by using thermometer probes specially fabricated for the present studies. Each thermometer probe consists of a plastic, bakelite or hylem tube of appropriate length with a calibrated thermistor element of high resistance temperature coefficient embedded at its tip. Two to four probes were used simultaneously to speed the survey as it takes time for them to attain steady state conditions.

The temperature probe is pushed into a hole bored to the desired depth by an auger or a hand operated steel rod of slightly larger diameter than the outer diameter of the probe. The holes are made of small diameter as these cause minimal disruptions to the thermal environment. Once the holes have been drilled, probes are inserted into these and left for some time to attain temperature stabilization.

Shallow temperature gradients have been measured by using a gradiometer specially designed and fabricated for these studies. The gradiometer consisted of five calibrated thermistors usually embedded in thin brass cylindrical pieces which were, in turn, fixed together at distances of 0, 20, 40, 60 and 80 cm, in thick walled bakelite tubes of outer diameter equal to 1.2 cm. The gradiometer after being inserted in holes were left for some time to attain equilibrium. The resistance values of thermistors were measured with the help of a resistance bridge, and a transistorized null detector.

Natural heat losses at the surface in certain parts of a hot spring area were also measured using a Heat Flow Transducer designed and pattented by the author (Indian patent No.85447). The heat flow transducer essentially consists of a thin thermopile having a large number of thermo junctions. The thermopile is calibrated so that its output in milli volts can be directly converted into heat flow values.

2.5 Survey Methodology for Exploration Of Geothermal Resources

Geothermal energy is defined as that amount of geoheat, which can be extracted and economically exploited. Areas where heat stored in rocks and water within the earth in shallow crustal layers ($z < 4$ km) is of magnitudes greater than the normal are economically viable for the development of geothermal resources. Naturally occurring water or steam or injected fluids can be made to transfer heat from such rock masses to the surface through a well. A combination of various techniques: Geological, Hydrogeological, Geophysical and Geochemical supported by drilling are used for exploration of geothermal energy resources.

Although natural thermal waters have been known to occur in a wide variety of geological environments, areas of potential geothermal resources are rather limited. In order to gain an insight into the geothermal resources of India and identify, if possible, certain large regions, where subsurface thermal waters may be encountered and exploited, an analysis and synthesis has been attempted of available geodata, particularly those relating to the major tectonic features of the Indian land mass. Certain geothermal provinces of India have thus been recognized.

2.5.1 Geophysical exploration

Geophysical exploration of anomalous subsurface features is based on the measurement and interpretation of physical fields perturbed by them or on anomalous emanations issuing from such regions. Their applications in delineating geothermal resources have been carefully examined and discussed in many review papers (Bodvarsson, 1970; Banwell, 1970, 73; Combs and Muffler, 1973; Singh and Gupta, 1977). Of the various geophysical techniques applicable to geothermal exploration the thermal and geoelectrical methods are the most widely used. A careful combination of these yields direct information about the nature, depth and size of a geothermal reservoir and also about the water movements in the geothermal system. Next in importance, are passive seismic techniques followed by self-potential, seismic, gravity and magnetic exploration, all of which provide quite useful though indirect information regarding the presence of a geothermal source.

Thermal and geoelectrical methods were used for exploration of geothermal resources for the first time in India in connection with the present studies and are discussed in a latër chapter.

Aquifers are geologic formations that contain and conduct water. They are found at depths ranging from a few metres to several kilometres. Confined aquifers are sandwiched between impermeable layers and saturated with water under pressure. Thermal aquifers have been exploited in many regions of the world for power generation, process heating and other municipal applications.

2.5.2 Geochemical thermometer

Several chemical geothermometers, both quantitative as well as qualitative, have been developed and used for evaluating the probable temperature of aquifers. Component concentrations of ratios mainly of Na, K, Ca and Silica, that can be related to subsurface temperatures are called the main quantitative geothermometers. These have been used to predict the probable reservoir temperatures of a large number of natural hydrothermal systems of India. The results are given in a later chapter.

The deep thermal waters do some time mix with the cold near surface waters. Various mixing models for warm and boiling waters have been proposed and used (Fournier and Truesdell, 1974; Gupta et al., 1976c). A warm mixing model, based on the observation of surface temperatures and the silica content of representative hot spring and cold waters was developed and used during the present study more or less at the same time when Fournier and Truesdell reportedly used it to predict probable reservoir temperatures. The probable residence time of hot waters in hot acquifers was inferred for some hydrothermal systems of India using Tritium measurements. The results are discussed in a later chapter.

2.5.3 Thermal methods for exploration of geothermal sources

Thermal exploration techniques exploit variations in the thermal conductivity between rock formations or in heat flow at deeper levels in different areas. Thermal anomalies are an expression of the subsurface thermal regime and their delineation can play a significant role in identifying potential geothermal resources.

2.5.3.1 Thermal anomaly in a geothermal area

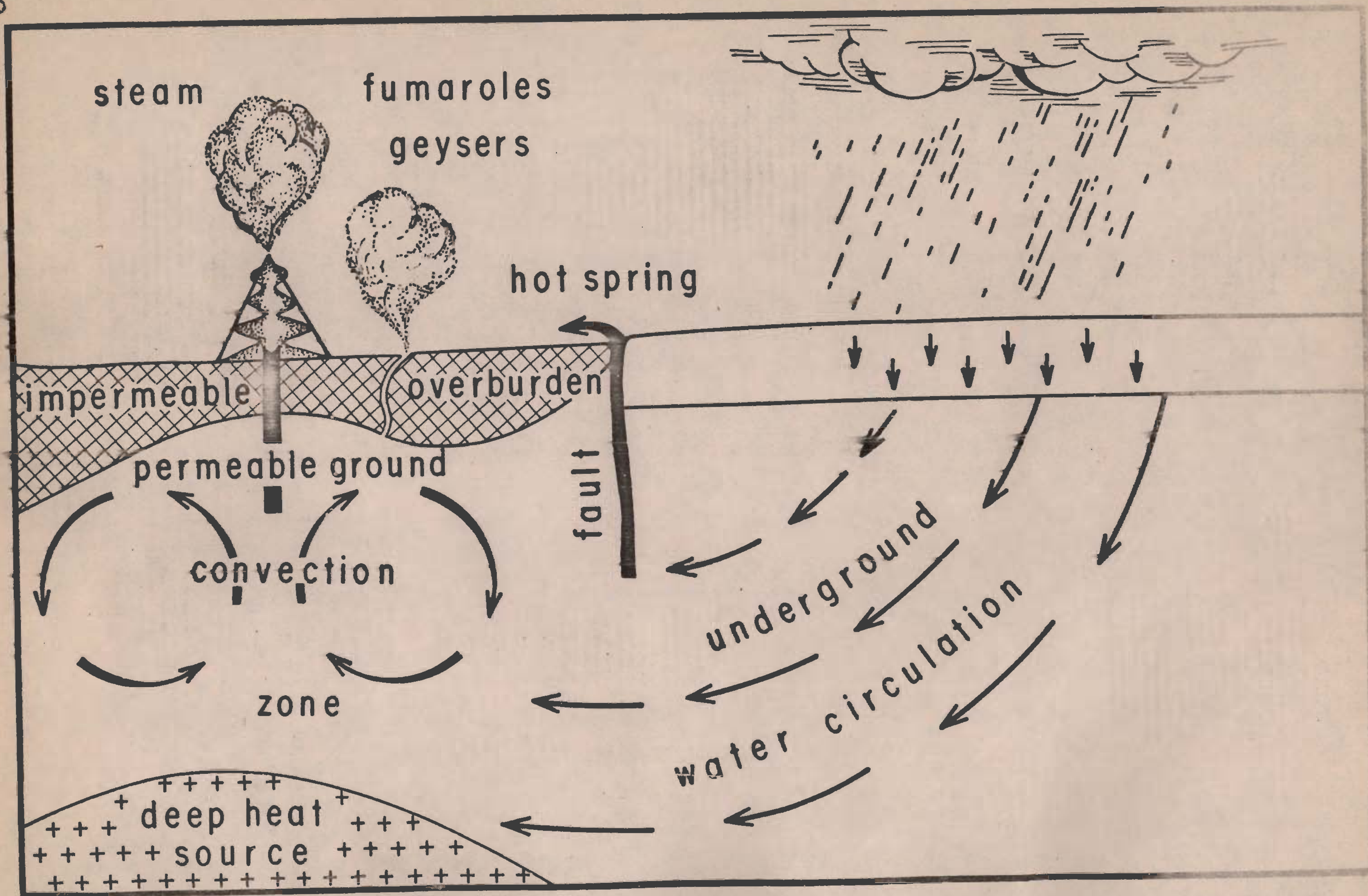
The main energy source of the earth resides in its hot interiors. Surface manifestations of this appear in the form of volcanoes, geysers, fumaroles and hot springs. Natural occurring hot water or steam existing in some suitable structures underground can be and has been tapped by drilling holes to appropriate depths. Such systems are known as natural hydrothermal systems. A great majority of known hydrothermal systems, that have been explored upto the present are associated with a large heat source at anomalously high temperatures present at depths of some thousands of metres. Porous and permeable formations overlying such anomalous heat zones and saturated with waters or steam, forming deep circulating systems (Figures II-5 and 6), exhibit strong thermal anomalies at the surface. Such systems are usually characterised by heat flows that are upto several thousand times above the normal (White, 1969) and therefore can be mapped by various thermal techniques which, in turn serve as direct means for assessing the areal extent and potential of geothermal reservoirs.

2.5.3.2 Thermal techniques for detecting thermal anomalies

Thermal techniques for exploration of geothermal energy resources can be divided into the following types:

1. Surface and near surface temperatures, and heat flow measurements, generally at depths of less than 5 m.
2. Geothermal gradient surveys, usually at depths of 30 to 100 m.
3. Heat flow investigations, usually in boreholes of depths more than 100 m; and
4. Air borne infra-red surveys.

Temperature and gradient surveys at 1-m depth, are


steam
fumaroles
geysers
hot spring
impermeable
overburden
permeable ground
fault
convection
zone
underground
water circulation
deep heat
source

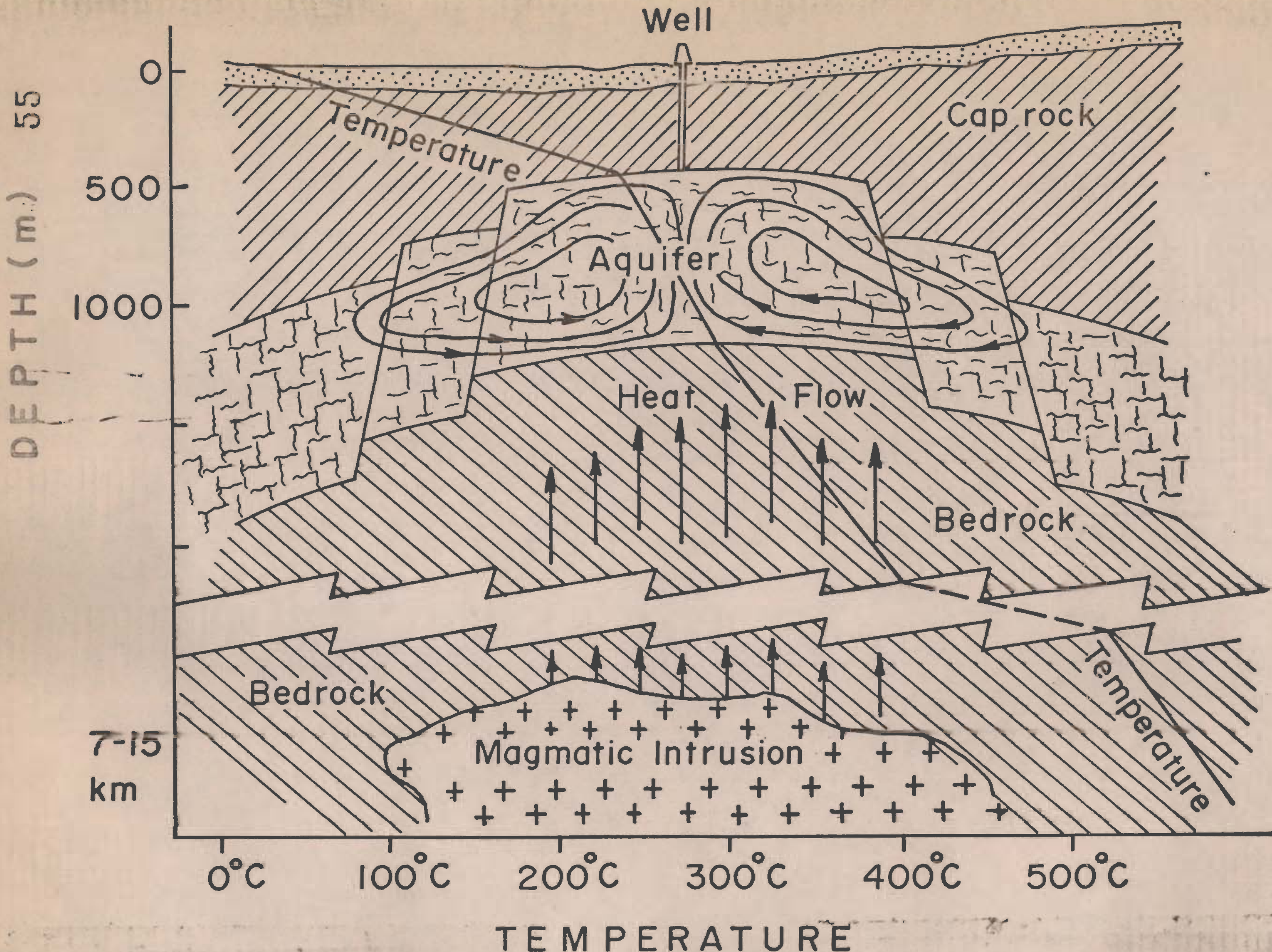


Fig.6 BASIC MODEL OF A GEOTHERMAL STEAM FIELD

quick and inexpensive and have been successfully used in hydrothermal areas as a rapid reconnaissance tool to detect, map and demarcate the boundaries of subsurface heat storage (Thompson, 1960; Robertson and Dawson, 1964; Gupta, 1967; Gupta and Rao, 1971; Gupta et al., 1974a,75b; Sharma and Gupta, 1976). The data obtained from such shallow depths often yield information about the circulation of thermal fluids in the shallow subsurface zones and sometime about the fissures or faults, and other disturbed zones in the subsurface which serve as channels for the hot ascending fluids (Gupta et al., 1975). The data can also be used to estimate the surface heat loss from a geothermal field which is an important part of the exploration programme as it enables one to assess the energy potential.

Temperature measurements at depths of 20-100 m are free from most of the near-surface thermal distrubances and yield reliable geothermal gradient data. However the data must always be evaluated for the effects of lateral movement of ground water. Over most of the potential geothermal areas, which are economically viable, temperature gradient at these depths are greater than 7°C/100 m, compared with the normal 3°C per 100 m. Such gradient measurements give the indication of the presence of a geothermal area. But, inorder to distinguish the main production zone from a large marginal zone, heat flow measurements have to be made in boreholes of depth more than 100 m (Combes and Muffler, 1973).

Thermal infra-red scanners are radiation detecting devices which scan the radiation emitted from each element of the scene. These normally operate in the 3 to 5-micron or 8 to 14-micron transmission windows in the atmosphere. Subterranean lava channels, geothermal resources, warm water channels in oceans which may incidentally be associated with mineral and water resources etc., can also be detected by the use of thermal scanners.

2.5.3.3 Detection of thermal anomalies at shallow depths

In order to be able to delineate thermal anomalies at shallow depths, one must consider the penetration of solar heat in the earth and the contribution of the earth's heat towards the heat budget of the shallow sub-surface regions.

Periodic variation of the Earth's surface temperature and its propagation:

Numerous meteorological observations have established that the air temperature at the earth's surface undergoes daily and annual variations which penetrate inside the earth with decreasing amplitude. In order to calculate the magnitude of the diuranal and annual temperature waves at different depths, we consider the fundamental Fourier heat conduction equation

$$\frac{\delta T}{\delta t} = \alpha \left[\frac{\delta^2 T}{\delta x^2} + \frac{\delta^2 T}{\delta y^2} + \frac{\delta^2 T}{\delta z^2}\right] \qquad \ldots \quad \text{II-12}$$

where α = thermal diffusivity = $K/\rho C$

$$\alpha = \frac{\text{thermal conductivity}}{\text{density x specific heat}}$$

For one dimensional flow,

$$\frac{\delta T}{\delta t} = \alpha \frac{\delta^2 T}{\delta x^2} \qquad \ldots \text{ II-13}$$

Let us consider the earth's surface to be a plane $z = 0$, and the surface temperature changes due to solar radiation as a sine function of time 't', then the oscillations of surface temperature around a mean temperature T_m are given by:

$$T_{o,t} = T_{o,a} \sin (2\pi nt) \qquad \ldots \text{ II-14}$$

Temperature excess ($T_{z,t}$), over the mean temperature values T_m, at any depth Z and at time 't' can be expressed by the equation (Jakob and Hawkins, 1957, p.68):

$$T_{z,t} = T_{o,a}\, e^{-Z\sqrt{\pi/\alpha P}} . \sin(\frac{2\pi t}{P} - Z\sqrt{\pi/P\alpha}) \quad \ldots \text{II-15}$$

and

$$\theta_z = T_m + T_{z,t} \qquad \ldots \text{ II-16}$$

where

θ_z = temperature calculated or observed at depth Z

$T_{o,a}$ = amplitude of the surface temperature wave

n = frequency, i.e., number of complete sinosodial cycles per unit of time

P = 1/n period of surface temperature wave, one day or one year

α = thermal diffusivity of the subsurface formations.

Expression II-15 is the solution of the heat conduction equation II-13 for the boundary condition as stipulated.

The amplitude of the Sine wave at depth Z becomes:

$$T_{z,a} = T_{o,a}\, e^{-Z\sqrt{\pi/P\alpha}} \qquad \ldots \quad \text{II-17}$$

It is clear that the amplitude rapidly decreases with depth.

There would be a lag between the times of the occurrence of a temperature maximum at the surface and at depth Z, the time lag being:

$$\Delta t = \frac{Z}{2}\sqrt{\frac{\pi}{P\alpha}} \qquad \ldots \quad \text{II-18}$$

For $Z = \sqrt{P\alpha\pi}$; then from equation (II-17)

$$T_{z,a} = T_{o,a}\, e^{-\pi} = \frac{T_{o,a}}{23}$$

i.e., the amplitude of the Sine wave is reduced to 1/23rd of its surface value at a depth $Z = \sqrt{P\alpha\pi}$

With α = 0.01 cm^2/sec; which is the average value for the surface rocks, and

a) Z = 52 cm for P = 24 hours
b) Z = 10 m for P = 365 days.

At twice the depths the amplitudes of the daily and annual temperature waves are reduced by a factor of

$$e^{-2\pi} = (1/23)^2 = 0.0019$$

The daily and annual temperature waves are not exactly of sinusoidal shape. These can be considered as a harmonic function of time following the sun, and represented by a Fourier series of the form (Carslaw and Jaeger, 1959, p.81):

$$\theta_{o,t} = T_m + \sum_{n=1}^{\infty} T_n \mathrm{Cos}(n\omega t - \epsilon_n) \quad \ldots \quad \text{II-19}$$

where

$\theta_{o,t}$ = temperature on the earth's surface at time 't'

T_n = amplitude of the partial waves (n^{th} harmonic)

ω = $2\pi/P$

T_m = mean temperature during the period (P), the period under consideration, e.g., one day or one year

ϵ_n = phase shift of the n^{th} harmonic.

If we consider the earth as a semi-infinite solid having surface temperature oscillations as given by equation II-19 then the temperature at depth 'Z' inside the earth at time 't' is given by

$$\theta_z = T_m + T_{z,t} \quad \ldots \quad \text{II-20}$$

where

$$T_{z,t} = \sum_{n=1}^{\infty} T_n \, e^{-Z\sqrt{n\omega/2\alpha}} \cdot \mathrm{Cos}(n\omega t - \epsilon_n - Z\sqrt{n\omega/2\alpha})$$

Equation II-20 is a damped wave equation giving

$$T_{z,a} = T_{o,a} \, e^{-Z\sqrt{\pi/P\alpha}}$$

Temperature at depth Z due to heat flow from the Earth's interior

Heat flows vertically towards the surface of the earth, and considering geological times, steady conditions have been obtained in most of the areas. In such a case:

$$q = K \frac{(T_z - T_s)}{Z} \quad \ldots \quad \text{II-21}$$

where,

T_s and T_z are temperature at the earth's surface and in the depth Z, respectively.

q = terrestrial heat flow

K = coefficient of thermal conductivity

As explained earlier 'q' has a very small magnitude. At the surface, heat is transferred to the environment, which acts as a large reservoir compared to 'q'. For such a condition:

$$q = A\,(T_s - T_a) \quad \ldots \quad \text{II-22}$$

where T_a is the air temperature near the surface and 'ZA' is the coefficient of surface heat transfer (cal./cm^{2}sec.°C).

Equating equations (II-21) and (II-22), one obtains

$$q = (T_s - T_a)A = (T_z - T_s).K/Z$$

$$T_z = q.Z/K + T_s$$

and

$$T_s = T_a + q/A$$

or

$$T_z = T_a + q/A + q.Z/K = T_a + \frac{q}{A}(1 + \frac{ZA}{K}) \ldots \quad \text{II-23}$$

If there are 'n' horizontal layers of thermal conductivities $K_1, K_2, \ldots \ldots K_n$ the equation (II-23) becomes

$$T_z = T_a + \frac{q}{A}\left(1 + A \sum_{1}^{n} \frac{Z_n}{K_n}\right) \qquad \ldots \text{ II-24}$$

Equation (II-23) gives the rise of temperature at a depth 'Z' due to heat flow from inside. Assuming the following values for A, q and K,

$A = 2 \times 10^{-4}$ cal.cm^{-2}.sec^{-1}.deg.C^{-1}

$q = 1.5 \times 10^{-6}$ cal.cm^{-2}.sec^{-1} (continental global average heat flow)

$K = 2 \times 10^{-3}$ cal.cm^{-1}.sec^{-1}.deg.C^{-1} (for partially saturated and loose surface material)

temperature rise due to terrestrial heat flow at depths of Z = 0, 0.8, 1, 2 and 20 metres are found to be respectively

$T_{0\,m} = 0.007°C$

$T_{0.8\,m} = 0.07°C$

$T_{1\,m} = 0.08°C$

$T_{2\,m} = 0.16°C$

$T_{20\,m} = 0.9°C$ (considering $K = 6 \times 10^{-3}$ cal./cm. sec.°C for $20 \geq Z > 2$ m).

It is therefore, clear that subsurface temperatures at shallow depths are controlled mainly by solar radiation, and contributions from the normal heat flow from the earth's interior are negligible.

If we assume the heat flow to be 5 times the normal value (≈ 7.5 HFU) then temperature anomalies of +0.4°C and 0.8°C would be observed at one and two metres depth respectively (vide equation II-23). However as mentioned earlier, in hydrothermal areas, strong positive thermal anomalies at the surface and in near-surface layers are caused due to convective heat and mass transfer, apart from heat flow due to conduction alone.

2.5.3.4 Estimation of natural heat loss

Various natural processes are responsible for the loss of natural heat on the surface in a geothermal area. Two ways to estimate this natural heat loss are: i) through estimation of the total heat which reaches the surface from below, and ii) through estimation of the total heat which is lost from the surface. Some time a combination of the two has to be used for a better estimate.

The flow of heat from the interior (Q_I) through a porous medium near the upper surface at which the thermal fluids are discharged occurs through: heat conduction (Q_1), convective fluid motion (Q_2), and mass flow (Q_3) i.e., heat escaping through fluid discharges in the form of hot springs, fumaroles and geysers, seepages. Adding all these heat losses

$$Q_I = Q_1 + Q_2 + Q_3 \qquad \text{... II-25}$$

Sometimes in geothermal areas, part of hot water which has been discharged on the surface and should have been measured is evaporated away, thus causing some loss of heat through evaporation.

Apart from heat escaping through mass flow from the surface (Q_{mf}), other processes which cause heat loss from the upper surface (Q_s) of hot ground are: simple heat conduction to the air from the surface, heat convection due to movement of air on the hot surface (Q_{con}), and evaporation from the wet hot surface i.e.,

$$Q_s = Q_{con} + Q_{conv} + Q_{ev} + Q_{mf} \qquad \text{... II-26}$$

Taking available emperical relations into considerations,

$$Q_s = \left[K_a \frac{d\theta}{dx}\right] \times A_1 + \left[(1 + 0.0069\Delta\theta) \times 2.04 \times 10^{-4}\right] \times A_2 + \left[(1+\frac{T}{150})(1+0.33W)(V-U) \times 0.33 \times 10^{-3}\right] \times A_3 + Q_{mf} \text{ cal./sec.} \qquad \text{... II-27}$$

where

$\Delta\theta$ = temperature difference between the surface and air

K_a = thermal conductivity of air

V = saturated vapour pressure of water at T°C

U = vapour pressure of air at mean temperature

W = air velocity

Q_{mf} = heat escape from the surface through mass flow in the form of hot springs, seepages to nearby bonds or streams or otherwise $\approx C \rho \sum_1 V_1 \Delta\theta_1$ cal/sec.

A_1, A_2, A_3 = respective areas in which heat loss is through the processes of conduction, convection and/or evaporation.

2.5.3.5 Heat flow from a geothermal reservoir

In a hydrothermal area, meteoric water from recharge areas percolates down from the surface and is heated up during its movement towards the reservoir (Figure II-6), where it remains for quite a long time at more or less a constant temperature. Subsequently, it reenters the soil to which it yields its heat. The annual mean temperature of the locality is thereby raised and strong thermal anomalies are caused in the vicinity of fissures, which can also be mapped by thermal surveys.

In order to compute the combined heat flow (Q) per unit time per unit area from a large reservoir, due to conductive, convective and mass transfer, we assume more or less steady conditions.

Let,

Z = depth of reservoir

A_o = the area of the anomalous zone on the surface

T_R = reservoir temperature °C

T_o = mean surface temperature outside the thermal zone

T_1 = mean surface temperature of the thermal zone

T_2 = temperature of the ground water table underlying the zone where geothermal fluids discharge on the surface

T_3 = temperature of hot fluid, when it enters the near surface layers after loosing its heat and mix with the cold water near the ground water table or spread over surface after its ascent from the reservoir

$w = w_1 + w_2 =$ quantity of hot water flowing out from the reservoir per unit time, cm^3/sec.

$w_1 =$ quantity of water reaching and mixing with cold water near the ground water table

$w_2 =$ quantity of water which directly reaches the surface and spreads over an Area A_1.

$w_3 =$ rate of ground water movement from the area cm^3/sec

$C =$ specific heat of water

$K =$ average thermal conductivity of the subsurface rock formations mcal/cm/sec °C

then,

$$Q = \frac{1}{A_o}\Big[\frac{K(T_R - T_2)}{Z} A_o + w_1.C.(T_3 - T_2) + w_2.C.(T_3 - T_1) + w_3.C.(T_2 - T_o)\Big] \quad \ldots \quad \text{II-28}$$

We have only considered a hot water system. If the reservoir is vapour dominated then the above equation has to be modified so as to account for the amount of heat carried through vapour and its condensation in the subsurface layers.

2.5.4 Geoelectrical methods

The usefulness and effectiveness of resistivity methods in delineating subsurface geothermal reservoirs has been well established in various geothermal fields of the world. At Larderello (Italy) the method was used more than two decades ago. A number of geothermal areas in the Taupo volcanic zone have been outlined using D.C. resistivity profiling (Hatherton, Macdonald and Thompson, 1966) as well as the Wairakai, Broad lands and Kaweran fields (Risk, Macdonald and Dawson, 1970; Macdonala and Muffler, 1972). D.C. Resistivity profiling using either Wenner or Schlumberger systems proved to be the most useful in the Taupo Graben area. Meidav and Banwell (1973) who analysed case histories of ten geothermal fields, concluded that almost without exception and regardless of the nature of host rocks, resistivities encountered in geothermal area were less than about 5.0 ohm.m, and that the presence of a steam phase within the reservoir should result in very high resistivities within the steam phase layer.

The resistivity of saturated rocks decreases with increasing temperature, effective porosity, electrolyte salinity and conductive mineral content. The rocks in a hydrothermal system can be expected to have very low resistivity. The method is found ideal for geothermal exploration, as resistivities within the geothermal areas ~~x~~ are one fifth to one twentieth of those outside. If the reservoir contains very little steam but has an abundance of hot water, it can prove to be an excellent target for exploration through resistivity methods. However, quantitative interpretations of VES curves become difficult in virgin areas on account of poor knowledge of subsurface temperature conditions, and of variations in the salinity of subsurface waters due to uncertain mixing of hot and cold waters. The effects of salinity and temperature are ~~thrm~~ therefore difficult to seperate on the VES cuRves.

2.5.4.1 Fundamental relationships

The relationship between the electrical resistivity of fluid saturated rocks and the various physical parameters, which affect it, have been discussed by numerous workers (Wyllie, 1963; Keller and Frischkencht, 1966; Dakhnov, 1962; Ward, and Fraser, 1967). Nevertheless, a brief review of the factors affecting current conduction in rocks are in order.

Current conduction in rocks other than shales, clays or metalliferous zones takes place mainly through groundwater in the pore spaces of rock bodies. According to Dakhnov (1962), the resistivity of a rock 'ρ_o' containing fluid of resistivity 'ρ_w' at 18°C is given by

$$\rho_o = \frac{\rho_w}{1 + \alpha(t-18)} = f(t)\ \rho_w \qquad \ldots \quad \text{II-29}$$

where,

α is the temperature coefficient of resistivity, usually 2.5 percent per degree centigrade.

Empirical data by Schlumberger Technology Corporation (1968) shows that logarithmic linear relationships exist between temperature and resistivity in the intermediate salinity range. At near-saturation concentrations, the effect of temperature on resistivity is somewhat non-linear.

It was empirically demonstrated by Archie (1942) that for different values of porosity in a given rock, there existed a constant relation, typical of that rock, between the resistivity of the rock, ρ_o, and the resistivity of the fluid saturating the given rock, ρ_w, given by:

$$\rho_o / \rho_w = F \qquad \text{... II-30}$$

where,

F, is known as the formation factor, a number that is characteristic of a rock for a given porosity.

This relationship was extended to relate the changes in the formation factor with those in porosity by equations of the type

$$F = k\,\phi^{-m} \qquad \text{... II-31}$$

where 'm' is a constant dependent upon the tortuosity of the mean free path of the electrical current through the rock matrix, often termed the cementation factor, and 'ϕ' is the fractional porosity. The value of 'm' is usually low (1.2 - 1.5) for well sorted, non-cemented sediments, and higher (1.9 - 2.2) for older well cemented or crystalline rocks; 'k' is a number near unity.

Combining equations (II-29), (II-30) and (II-31)

$$\rho_o = K\,\phi^{-m}\,\rho_w\,f(t) = F\,\rho_w\,f(t) \qquad \text{... II-32}$$

Equations (II-32) state that the true resistivity of the rock is a function of porosity, tortuosity, salinity of the saturating electrolyte and temperature of the electrolyte. It has been impilcitly assumed above that there exists no solid conduction, such as due to clays or metallic minerals. The effect of solid conduction through the matrix is to decrease the overall resistivity of the rock, ρ_o, resulting in an apparent formation factor which is always lower in value than the true formation factor.

2.5.5 Geothermometry

Geothermal fluid chemistry constitutes a powerful tool in geothermal exploration. Geochemical investigations of hot springs, fumaroles and cold surface waters can supply critical informations regarding the characteristic of the source region, path of fluid movement and the turnover time of water in the system. It also exposes the nature of the hydrothermal system: steam heated or hot water, subsurface reservoir temperature, extent of mixing of deep thermal waters with shallow cold waters, zone of high upflow, and permeability.

Isotopic compositions of most of the high-temperature waters have indicated that their origin is dominantly meteoric (Craig, 1963). However, some hot fossil brine and probable hot metamorphic water (White et al. 1973), have also been reported. The underground percolating meteoric water is affected with two main complimentary physico-chemical factors; the subsurface temperatures and the interaction with wall-rock minerals. The concentrations in the natural water of the common rock forming minerals such as silica, aluminum, sodium, potassium, calcium, magnesium, iron, and manganese are controlled by the particular temperature dependent - mineral - water - equilibria. The solubilities of the main rock forming minerals and the existence of temperature-dependent equilibria at depth has helped to establish qualitative as well as quantitative, geochemical indicators of subsurface reservoir temperatures. The quantitative type of geochemical thermometers are described below.

The main quantitative chemical indicators of temperature which have attracted most attention are the Silica and Na/K ratio geothermometers. Recently, Fournier and Truesdell (1973) have developed another namely the Na-K-Ca geothermometer. The first thermometer is based on the silica content of thermal waters. In hydrothermal areas, silica at depth, occurs in various forms: quartz, chalcedony, cristobalite and amorphous silica. Quartz seems to control the solubility of silica in deep waters from high temperature (150°C) geothermal areas. Close correlation between temperatures directly measured in geothermal wells (180° - 260°C) and those estimated from measured silica concentrations assuming equilibrium with quartz has been reported by Mahon (1966), Fournier and Rowe (1966) and Arnorsson (1975). According to Fournier and Truesdell (1970) lower temperature waters may be saturated with chalcedony rather than quartz. Arnorsson (1975) found that Icelandic geothermal well waters at temperatures below 110°C agreed with chalcedony solubility. The temperature of deep waters in equilibrium with quartz can be estimated using the following formula (Truesdell, 1975).

$$t°C = 1315/(5.205 - \log SiO_2) - 273.15 \quad \ldots \quad \text{II-33}$$

where Silica concentration is in ppm. The above equation is valid when water from an aquifer moves to the surface slowly enough for conductive cooling to occur. For adiabatic isoenthalphic cooling, the formula is:

$$t°C = 1533.5/(5.768 - \log SiO_2) - 273.15 \quad \ldots \quad \text{II-34}$$

The above equations give good results in the temperature range of 125° to 225°C.

For probable conditions at lower temperatures, for chalcedony conductive cooling, Truesdell (1975) has given the following formula.

$$t°C = 1015.1/(4.655 - \log SiO_2) - 273.15 \quad \ldots \quad \text{II-35}$$

Next to SiO_2, the Na/K ratio of the thermal water is a significant indicator of the subsurface aquifer temperature (White, 1965; Ellis and Mahon, 1967). This thermometer is based on the postulation (White, 1965) that Na/K ratios in most thermal waters are controlled by equilibrium with albite and K-feldspar and the assumptions that the activities of the solid species are unity and that activity coefficients for Na and K cancel each other, so that forthe exchange relation?

$$K^+ + Na^+ - \text{feldspar} = \text{K-feldspar} + Na^+ \quad \ldots \quad \text{II-36}$$

the equilibrium constant K_e = molar Na/K

For such a case the Van't Hoff equation, which gives the variation of K_e with temperature, can be used for the prediction of reservoir temperatures by using Na/K ratios. The Van't Hoff equation is given by the following relation.

$$\frac{d \log K_e}{d(1/T)} = - \frac{H^\circ(T)}{4.5758} \qquad \ldots \text{ II-37}$$

where,

T = absolute temperature

$H^\circ(T)$ = standard hear of reaction at given temperature.

K_e = the equilibrium constant = molar Na^+/K^+

It is well known that $H^\circ(T)$ of the above mentioned exchange reaction has a very small variation with temperature. Using this criterion and synthesizing the available data from high temperature geothermal areas, White (1970). Ellis (1970) and Fournier and Truesdell (1973) gave more or less straight line curves between Na/K atomic ratios and $1/T^\circ K$. White-Ellis curve is the more widely used alongwith the following relation given by Truesdell (1975) for their curve.

$$t^\circ C = 855.6/(\log Na/K + 0.8573) - 273.15 \quad \ldots \text{ II-38}$$

where Na/K concentrations are in ppm.

The Na/K geothermometer has been successfully used for the prediction of reservoir temperatures for near neutral pH geothermal waters, which do not deposit travertine and have values of $Ca^{\frac{1}{2}}/Na$ less than unity (for molal concentrations).

It has been generally observed that the Na/K geothermometer gives anomalously high temperatures for waters high in calcium and does not work at temperatures below 100 - 120°C. Considering the influence of calcium in aluminosilicate telations, Fournier and Truesdell (1973) proposed the following empirical Na-K-Ca geothemometer for such waters.

$$\log_{10}Na/K + \beta \log_{10}(Ca^{\frac{1}{2}}/Na) = \frac{1647}{T^{\circ}(K)} - 2.24 \quad \ldots \text{ II-39}$$

or

$$t^{\circ}C = \frac{1647}{\log_{10}(Na/K) + \beta \log_{10}(Ca^{\frac{1}{2}}/Na) + 2.24} - 273.15 \quad \ldots \text{ II-40}$$

The value of β is chosen depending upon whether the equilibrium of water occurs above or below 100°C, and

$$\beta = 4/3 \text{ for } Ca^{\frac{1}{2}}/Na > 1, \text{ and } t \leq 100^{\circ}C \quad \ldots \text{ II-41}$$

$$\beta = 1/3 \text{ for } Ca^{\frac{1}{2}}/Na < 1, \text{ or } t \text{ for } \beta = 4/3 > 100^{\circ}C \quad \ldots \text{ II-42}$$

where Na, K, Ca concentrations are in moles/litre.

The upward moving natural thermal waters often change compositionally during transit by boiling and dilution and may sometime be even contaminated with brine at ~~[illegible]~~ depth.

Na-K and Na-Ca-K geothermometers are based on the ratios of the concentration of these elements in water. The surface or shallow cold waters mostly have very low concentrations of these constituents. It can be easily shown that mixing of highly saline deep waters with fresh cold waters of very low salinity will not greatly change the magnitude of the ratios of these elements and does not therefore appreciably affect the estimated reservoir temperatures. The limitations of the aforestated chemical geothermometers have been discussed in detail by Gupta and Saxena (1979).

CHAPTER III

HEAT FLOW IN CAMBAY BASIN AND AT ALORE-KOYNA (DECCAN TRAPS)

3.1 Introduction

Thermal measurements were made in a number of wells in different fields of the Cambay basin, which were initially drilled by the ONGC for oil (vide Table III-1) and later stabilized by displacing the mud by water. Actual measurements were made after 6 months of the wells having been thus prepared using a special thermistor probe designed and fabricated for this purpose. Results obtained in the Cambay, Kathana, Nawagam, Sanand, Kalol, Mehsana, Ankleswar and Broach oil and gas fields of the Cambay basin, as well as heat flow values obtained in the Alore-Koyna region are discussed in this chapter.

3.2 Geology Structure and Tectonics of the Cambay Basin

The Cambay basin is located in the alluvial plains of the Gujarat State of Western India, approximately bounded by the latitudes 21°N and 24°N and longitudes 71°30' and 73°45'E. It is a Cenozoic basin running as a narrow graben in an approximately north-northwest south-southeast direction uptil latitude 21°45' where it swings towards NNE-SSW and runs into the Gulf of Cambay. Extensive geophysical work in connection with oil exploration has been carried out in the

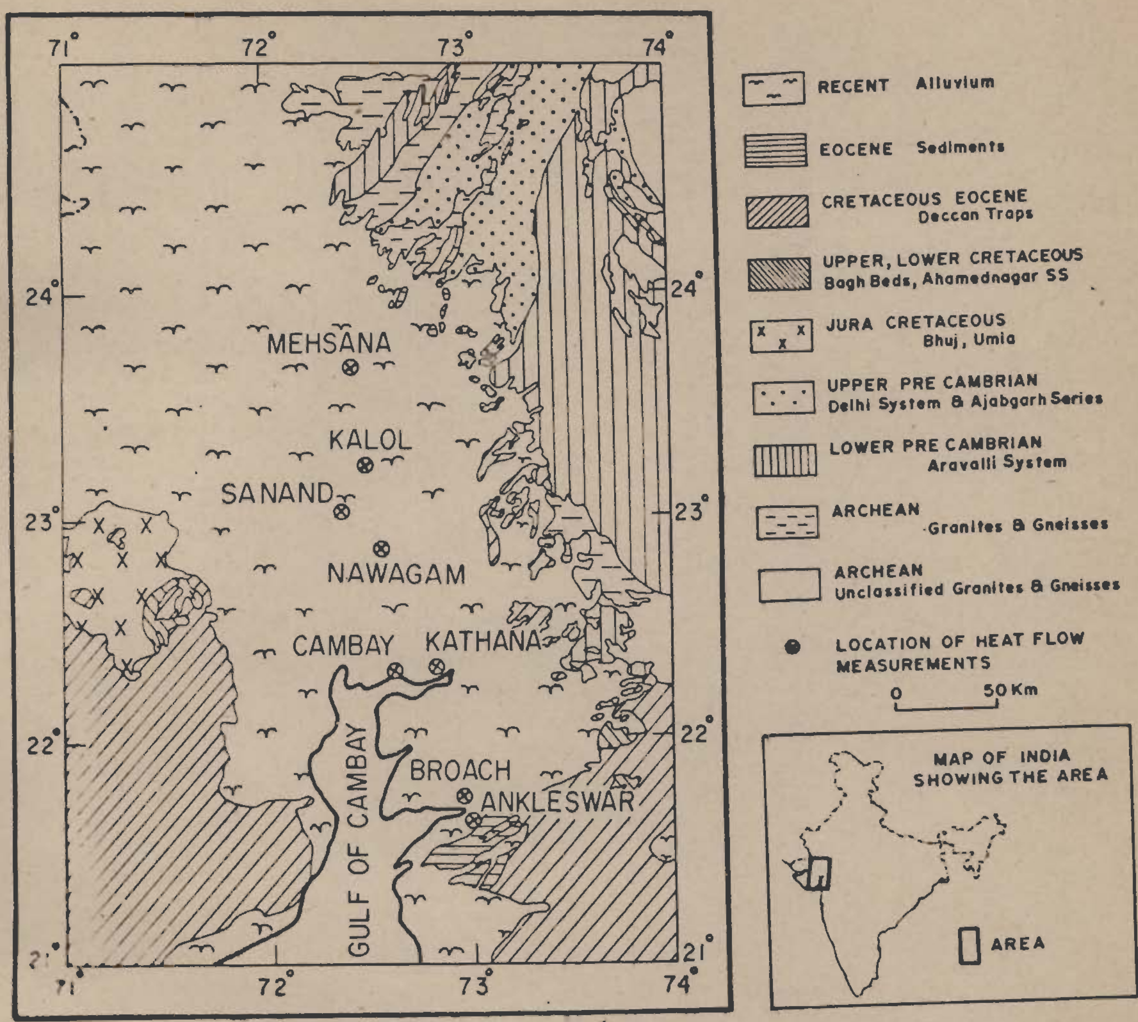


FIGURE III-1

Geology of Cambay Basin and Adjacent Region (Geology from Geological Survey Map of India, 1963).

basin since 1958 and its geology and tectonics discussed in detail by many workers including Mathur et al. (1968); Sen Gupta (1967); and Mathur and Evans (1964).

Tectonically, the basin is situated at the northwestern edge of the Peninsular Shield. It is bounded by the Aravalli System of Precambrian age on the northeast, and by the Deccan Traps on the east and west. The geological map of the basin and adjoining regions is shown in Figure III-1.

A complete sequence of sediments ranging in age from Recent to Eocene overlie an irregular surface of Deccan Traps. The contact is however exposed only at a few places and appear to be normal except southeast of the river Narmada. The sequence comprises greywackes, dark grey to black grey shales, coal cyclothems, silts, fine to medium grained sands, and grey reddish-brown clays. The sediments reach a thickness of over 5,000 m in the deepest part of the basin i.e., the Jambusar-Broach area (Figure III-2). The presence of sporadic outcrops of Jurassic and Cretaceous sediments on its margins points to the likelihood of this basin having existed in the Mesozoic era.

It is believed that a deep-seated fault, probably formed towards the end of the Gondwana Period, runs parallel to the western coast of India and into the Cambay basin. This fault zone is cut into two parts by an important transverse basement fault running along the Narmada River

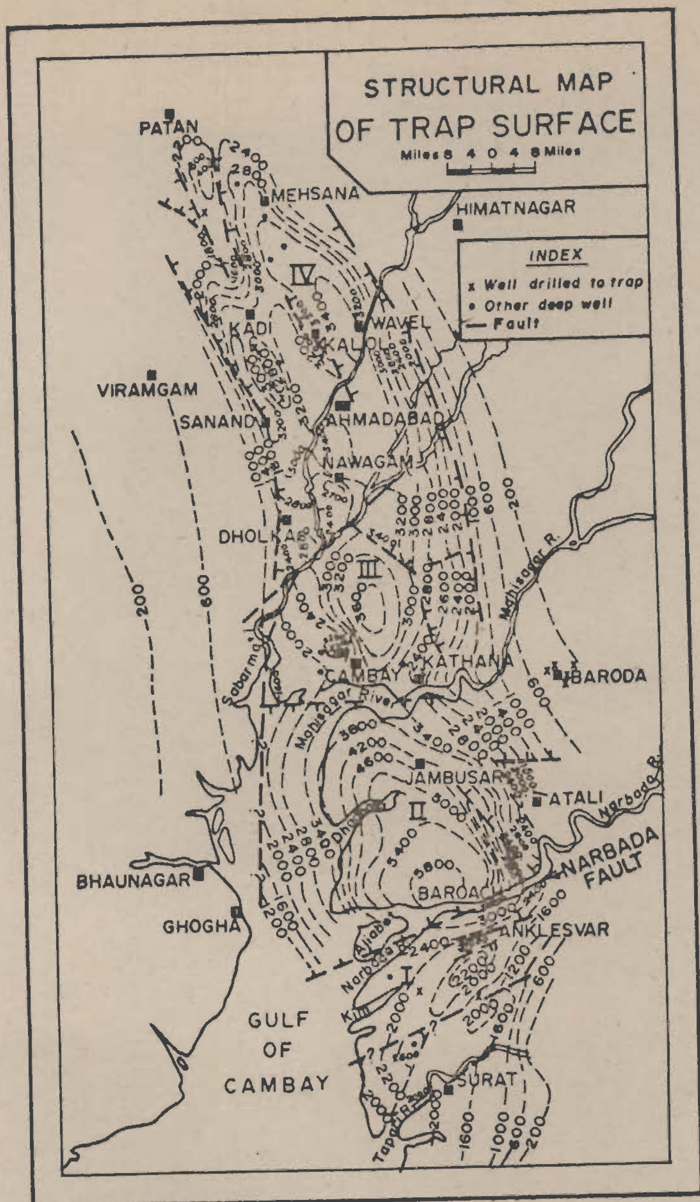


FIGURE III-2

Map of the Trap Surface in the Cambay Basin With Four Tectonic Blocks (I-IV) After Avasthi et al. (1969)

and joining up with the southern coast of the Saurashtra Peninsula. The area to the south of the Narmada River forms the northeastern corner of the southern part of the fault zone. The other part of the fault zone comprises the rest of the Cambay basin (Mathur et al., 1968).

The Narmada valley is an old zone of weakness, movements along which have occurred at widely different periods. South of the Narmada, the folding in the Cambay basin shows a strong ENE-WSW trend, roughly parallel to the Narmada fault. In the northern part upto as far as Mahisagar (Figure III-2) there are few secondary structural features and no specific trend is discernible. North of Mahisagar, the structural trends are NNW-SSE or N-S; apparently related to the axial direction of the basin (Mathur and Evans, 1964).

The whole basin is dissected into four structural blocks (Figure III-2) by faults within the Deccan Traps, which somewhat extent into the overlying sediments. These blocks are characterised by different fold and fault trends and basement depths, the sedimentation being partly controlled by the block pattern.

Another conspicuous feature of the basin is the reversals of block tilting throughout its Cenozoic history. During the Eocene, the general slope of the basin was towards the north, but it was reversed during the Oligocene, and the basin as a whole became tilted towards the south. The tilting of the basin towards the south continued during the Miocene Epoch. The Narmada fault, which was dormant during the major part of the Miocene Epoch, was reactivated during the post-Miocene period, and the movement along this fault resulted in a vast syntectonic accumulation of sediments in the Jambusar-Broach area and in the uplift and consequent erosion of the Narmada block to the south. The southerly tilt of the basin continued during the post-Miocene. A slight but progressive westerly shift in the axis of sedimentation during Cenozoic history is noticeable.

3.3 Temperature Measurements

Temperature measurements were made in 15 oil wells, three each at Ankleswar, Cambay and Kathana, two each at Nawagam and Kalol, one each in the Sanand and Mehsana oil fields. Measurements were made at depth intervals of 20 or 25 m, upto a maximum depth of 1,250 m. The locations and other particulars of these wells are given in Table III-1. Bottom hole temperatures as measured by ONGC in exploratory well No. Broach-6 have been used also for calculation of heat flow for the area.

Least square fitted temperature gradients were calculated for those depth intervals within which the temperature - depth curves were found to be substantially linear and are given in Table III-2. Temperature gradients were found to be lowest in Broach and highest in the Cambay-Kathana-Nawagam region as compared with other parts of the basin.

TABLE III-1

Particulars of Wells Logged in Cambay Basin and the Heat Flow Data

Locality & Coordinates	Well No.	Elevation (m) above m.s.l	Depth to which logged (M)	Undisturbed period (month)	Heat flow u gal/ cm^{2}sec.
Mehsana Oil field 30°36'N, 72°11'E	M-6	60.7	1120	10	1.9
Kalol Oil field 23°16'N, 72°30'E	K-58 K-27	73.8 65.2	1200 1220	6 30	1.8 1.9
Sanand Oil field	S-5	41.4	1060	66	1.8
Nawagam Oil field 22°50'N, 72°30'E	Naw-32 Naw-43	-- 24.2	900 990	-- 18	1.9 2.0
Kathana Oil field 22°17'N, 72°48'E	Kat-9 Kat-5 Kat-3	26.2 20.7 --	1200 1200 900	7 12 12	2.2 2.0 2.2
Cambay Gas field 22°33'N, 72°35'E	C-10 C-36 C-33	11.9 12.2 15.7	1225 1100 1200	60 50 50	2.4 2.3 2.0
Ankleswar Oil field 21°35'N, 72°55'E	Ank-2 Ank-28 Ank-170	12.5 13.0 23.3	1200 1200 1200	70 50 19	1.7 1.65 1.5
Broach	6	--	1700*	--	1.3*

*from bottom hole temperature data

TABLE III-2

Geothermal Gradients in Various Wells in Cambay Basin.

Well No.	Depth interval (m)	Temperature gradient (°C/km)
Mehsana-6	600- 910	41.1±0.7
	980-1120	59.2±0.5
Sanand-5	600- 840	56.4±1.3
	840- 900	40.8±0.9
	900-1050	41.6±1.4
Kalol-58	560- 675	37.2±2.3
	675- 875	48.4±0.8
	875- 975	31.1±3.4
	975-1075	40.0±2.0
	1075-1200	59.1±2.0
Kalol-27	560- 740	39.1±1.5
	740- 900	48.7±1.3
	920-1020	30.1±2.2
	1020-1140	39.7±0.8
	1140-1220	59.0±5.7
Nawagam-32	400- 525	43.3±2.1
	550- 800	61.8±1.0
	800- 900	48.7±1.2
Nawagam-43	600- 860	66.5±1.4
	900- 980	38.2±1.9
Kathana-9	500- 775	48.4±0.7
	775- 900	51.3±2.7
	900-1050	63.9±2.0
	1075-1150	70.7±2.0
Kathana-5	325- 550	42.7±0.7
	550- 825	46.5±0.4
	850- 925	50.8±1.4
	925-1075	64.3±1.0
	1075-1200	54.2±1.0
Kathana-3	350- 475	40.0±0.8
	500- 600	52.9±0.5
	625- 850	45.0±1.0
Cambay-10	450- 600	53.3±0.7
	625- 775	48.8±1.0
	875- 975	74.9±8.0
	975-1125	65.7±1.7
Cambay-36	450- 625	50.0±0.8
	625- 750	44.5±1.8
	775- 850	46.9±6.0
	850-1100	63.9±0.8
Cambay-33	525- 700	46.8±0.6
	700- 850	42.5±0.9
	1025-1175	52.2±1.3
Broach-6	500-1220	26.9±1.1
Ankleswar-2	750-1200	47.3±0.3
Ankleswar-28	350- 900	41.9±0.9
	900-1075	53.1±1.9
	1075-1175	38.9±3.0
Ankleswar-170	525- 675	32.2±1.1
	675- 800	33.5±2.2
	800-1125	41.6±1.9
Ankleswar-5	775- 850	29.3±0.5
	850-1150	43.0±1.2

3.4 Measurements of the Coefficient of Thermal Conductivity

Accurate estimation of the mean conductivity coefficient of stratigraphic columns comprised of varied sequences of sediments at various depths in a drill hole, poses a major problem and very few of these values in respect of claystones, shales, and silty shale are available in published literature.

Unfortunately no core samples could be obtained from the wells which were actually investigated, but some representing major rock types of the Cambay basin were obtained from various other wells in the basin notably from Kalol, Cambay and Ankleswar. Circular discs of 41.2 mm dia. were used for determination of the coefficient of thermal conductivity on the Modified Birch's apparatus using quartz discs cut parallel to the optic axis as a standard. The conductivities were first determined in an over-dry state and then after saturation with Kerosene under vacuum to prevent disintigration. For the few samples which did not disintegrate in water, the conductivity values were also determined after saturating them with water under vacuum. The values obtained for various types of sediments are given in Table III-3.

Conductivity coefficients (K_{sw}) of water saturated samples of claystone, shale and silty sandy claystone and shale, could not be measured as they disintegrated rather

TABLE III-3

Thermal Conductivity of Core Samples
From Cambay Basin

Sample No.	Depth from surface(m)	Rock type	Thermal Conductivity mcal/cm. sec. °C Air saturated (K_{sa})	Oil saturated (K_{so})	Water saturated (K_{sw})	Ratio $\frac{K_{so}}{K_{sa}}$	$\frac{K_{sw}}{K_{sa}}$
1	2	3	4	5	6	7	8
C-7	1000	Clay Stone	1.78	2.31	--	1.29	--
C-8	698	Clay Stone	1.93	2.50	--	1.29	--
C-9	798	Clay Stone	1.75	2.47	--	1.40	--
C-13	801	Clay Stone	2.30	2.84	--	1.24	--
C-20	703	Clay Stone	2.04	2-53	--	1.23	--
C-1	--	Clay Stone	1.87	--	--	--	--
A-11	453	Clay Stone	2.57	3.48	--	1.35	--
C-23	1505	Shale	1.71	2.25	--	1.31	--
C-2	1176	Clay Shale	2.13	2.85	--	1.34	--
K-9	1070	Shale	2.54	3.16	--	1.24	--
K-12	1478	Shale	1.85	2.41	--	1.30	--
K-6	984	Shale	1.81	2.16	2.89	1.19	1.59
A-5	926	Shale	2.28	2.80	--	1.22	--
A-12	850	Clayey Shale	1.99	2.52	--	1.26	--
A-6	910	Clayey Shale	2.22	2.59	3.42	1.16	1.32
A-7A	1056	Shale	1.72	--	--	--	--
A-7B	1056	Shale	1.98	2.36	--	1.20	--
C-19	1305	Silty Sandy Shale	2.66	3.28	--	1.23	--
C-22	1175	Silty Clay Stone	--	2.71	--	--	--

Table III-3 (contd.)

1	2	3	4	5	6	7	8
K-13A	1044	Silty Shale	2.99	--	--	--	--
K-13B	1044	Silty Shale	2.96	4.08	--	1.38	--
K-11	1200	Silty Shale	3.37	- -	--	--	--
K-1	1055	Silty Shale	3.24	3.71	--	1.14	--
K-14	1040	Silty Shale	2.78	3.84	--	1.39	--
K-14	1040	Silty Shale	3.17	--	--	--	---
K-8	695	Silty Clay	2.46	3.00	--	1.22	--
K-18	1343	Silty Clay	2.86	3.48	--	1.22	--
K-8	900	Silty Clay	2.44	2.89	--	1.18	--
K-7	1050	Silty Clay	2.48	--	--	--	--
K-16A	2803	Sand Stone	3.95	4.67	6.25	1.18	1.58
K-16B	2806	Sand Stone	4.24	4.86	6.28	1.15	1.48
K-4	900	Sand Stone	4.12	4.60	6.45	1.15	1.56
K-2	929	Mottled Sandy siltstone	4.28	4.69	6.54	1.09	1.53
C-24	1295	Lime Stone	--	--	5.21	--	--
A-11	1136	Lime Stone	3.62	4.20	5.34	1.16	1.47
A-3	670	Fossiliferous lime stone	2.85	3.23	4.90	1.13	1.72
A-14	807	Silt Stone	2.52	3.30	5.02	1.31	1.99
A- 13	1007	Silt Stone	3.12	3.89	5.50	1.25	1.76
C-14	1298	Silt Stone	3.17	4.08	5.54	1.29	1.75
K-2	930	Silt Stone	4.03	4.41	6.50	1.09	1.61

quickly in water. The ratio of 'water saturated' conductivity (K_{sw}) to 'dry' conductivity (K_{sa}), which could only be obtained for a few samples was found to vary from 1.32 to 1.59 for shales, clayey shales and sandstones, & with an average of about 1.50. The ratio is much higher for samples of fossiliferous limestones and the siltstones. A value of 1.50 for K_{sw}/K_{sa} was, however, adopted for estimating the 'water saturated' conductivity values finally adopted for various rock types are given in Table III-4. For clayey and shaley sandstones, which are a few metres thick in the stratigraphic columns, a value of 5.5 ± 0.3 mcal/cm sec.°C, which is intermediate between those for sandstones and silty shales was adopted. No representative samples of this rock type could be obtained for conductivity studies.

3.5 Evaluation of Heat Flow

The heat flow values were calculated, vide section 2.3, after first correcting the water saturated conductivity values for temperature increase with depth as suggested by Joyner (1960).

3.5.1 The Cambay oil and gas field

The Cambay (or Lunej) field 9 km NNW of the town of Cambay and about 60 km west of Baroda, was the first commercially productive field to be discovered in the Cambay basin. The field lies over a north-south trending anticline faulted on its east flank and on its west by minor faults

TABLE III-4

Summary of the Conductivity Values Measures and Adopted

Rock type	No. of samples	Mean conductivity (mcal./cm.sec.°C) Air saturated K_{sa}	Oil saturated K_{so}	Water saturated conductivity values adopted
Clay stones	8	$2.03 \pm 0.1^{*}$	$2.69 \pm 0.2^{*}$	$3.0 \pm 0.2^{*}$
Shales and Clayey shales	10	2.02 ± 0.1	2.57 ± 0.1	3.0 ± 0.2
Silty or sandy shales or silty sandy claystones from Kalol oil field.	8	2.98 ± 0.29	3.62 ± 0.40	4.5 ± 0.45
Lime Stone	3	3.23 ± 0.54	3.71 ± 0.68	5.2 ± 0.23
Silt Stone	4	3.21 ± 0.62	3.92 ± 0.46	5.6 ± 0.62
Clayey or Shaly sand stones	-	-	-	5.5 ± 0.3

*standard error of the mean

(Mathur and Evans, 1964). A close examination of the temperature depth curves (Figure III-3) of the 3 wells, C-10, C-36, and C-33 of the Cambay field show specific depth intervals in which the temperature depth curves are fairly linear. The heat flow values have been calculated for these eleven depth intervals using weighted mean conductivity values and least square fitted gradients for the specific depth interval. These (Table III-5) show considerable variation and distinguish C-10 from C-36 on one side and C-10 from C-33 on the other. Temperature (Figure III-3) are also lower for C-33 as compared with those at corresponding depths in C-10 and C-36. Wells C-10 and C-36 are located near the top of the Lunej structure, while well No.C-33 is located on its eastern flank.

3.5.2 The Kathana oil field

This field lies over a NE-SW trending anticline and is situated east of the famous Cambay gas field. Heat flow values were determined in three well Nos. Kat-3, 5 and 9 using the BP technique. An increase of heat flow values was noted at a depths of about 650, 725 m in well Nos. Kat-3 and 9(Table III-6). The temperature profile, the percentage of rock formations and the Bullard plot for well Kat-9 are shown in Figure III-4.

3.5.3 The Nawagam oil field

The Nawagam anticline, which lies about 24 km south of Ahmadabad, is irregular in shape, but has a general east-

TABLE III-5

Mean Thermal Conductivity, Gradient and Heat Flow for Three Cambay Wells

Well No.	Depth interval (m)	Gradient (°C/km)	Weighted mean conductivity in the depth interval (mcal/cm Sec°C)	Heat Flow (μ cal/cm^{2}sec)
C-10	450- 600	53.3±0.7*	3.4±0.3*	1.8±0.2*
C-10	625- 775	48.8±1.0	3.6±0.3	1.8±0.2
C-10	875- 975	74.9±8.0	3.3±0.3	2.5±0.3
C-10	975-1125	65.7±1.7	3.5±0.2	2.3±0.2
C-36	450- 625	50.0±0.8	3.6±0.3	1.8±0.2
C-36	625- 750	44.5±1.8	3.9±0.2	1.7±0.2
C-36	775- 850	46.5±6.0	3.8±0.3	1.8±0.3
C-36	850-1100	63.9±0.8	3.6±0.2	2.3±0.1
C-33	525- 700	46.8±0.6	3.4±0.3	1.6±0.1
C-33	700- 850	42.5±0.9	4.1±0.3	1.8±0.1
C-33	1025-1175	52.2±1.3	3.9±0.2	2.0±0.1

*the limits are given at 95% confidence level

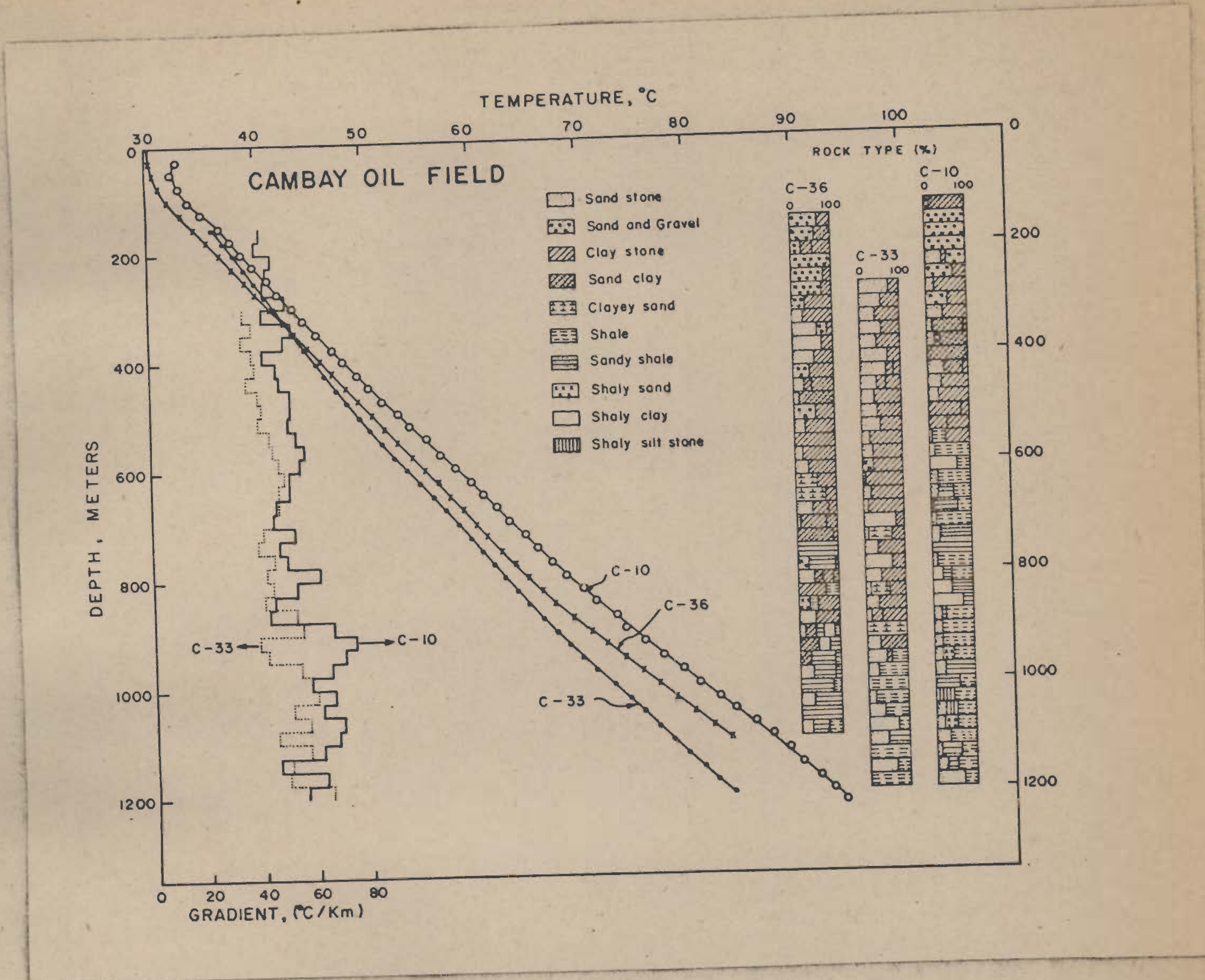


FIGURE III-3

Temperature vs Depth Curves, Stratigraphy and gradients in Cambay Wells

TABLE III-6

Heat Flow Data from Different Depth Intervals for Mehsana, Kalol, Sanand, Nawagam, Kathana, Ankleswar and Broach Wells

Locality	Well No.	Depth inter-val(m)	Heat flow (u cal/cm^2 sec.)
Mehsana Oil field	6	900-1120	1.9
Kalol Oil field	K-58	550- 975 975-1200	1.7 1.8
	K-27	560-1000 1020-1220	1.8 1.9
Sanand Oil field	S-5	800-1000	1.8
Nawagam Oil field	32	400- 900	1.9
	43	600- 860	2.0
Kathana Oil field	Kat-9	525- 725 725-1200	1.9 2.2
	Kat-5	700- 900 900-1200	2.0 2.0
	Kat-3	500- 650 650- 900	1.9 2.2
Ankleswar Oil field	A-2	750-1200	1.7
	A-28	400- 500 500-1175	1.5 1.65
	A-170	700-1200	1.5
Broach	6	420-1220	1.3

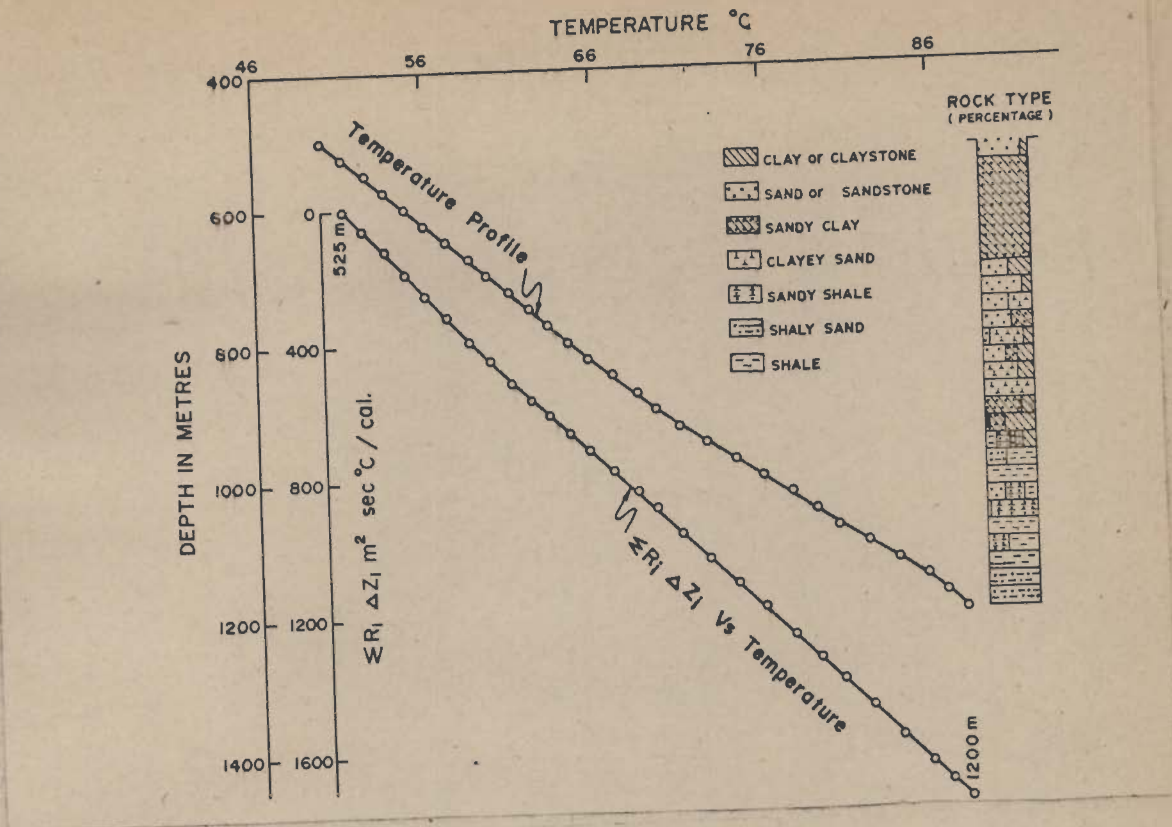


FIGURE III-4

Temperature vs Depth Curve, Lithology and Bullard Plot for Well No.9, Kathana Oil Field.

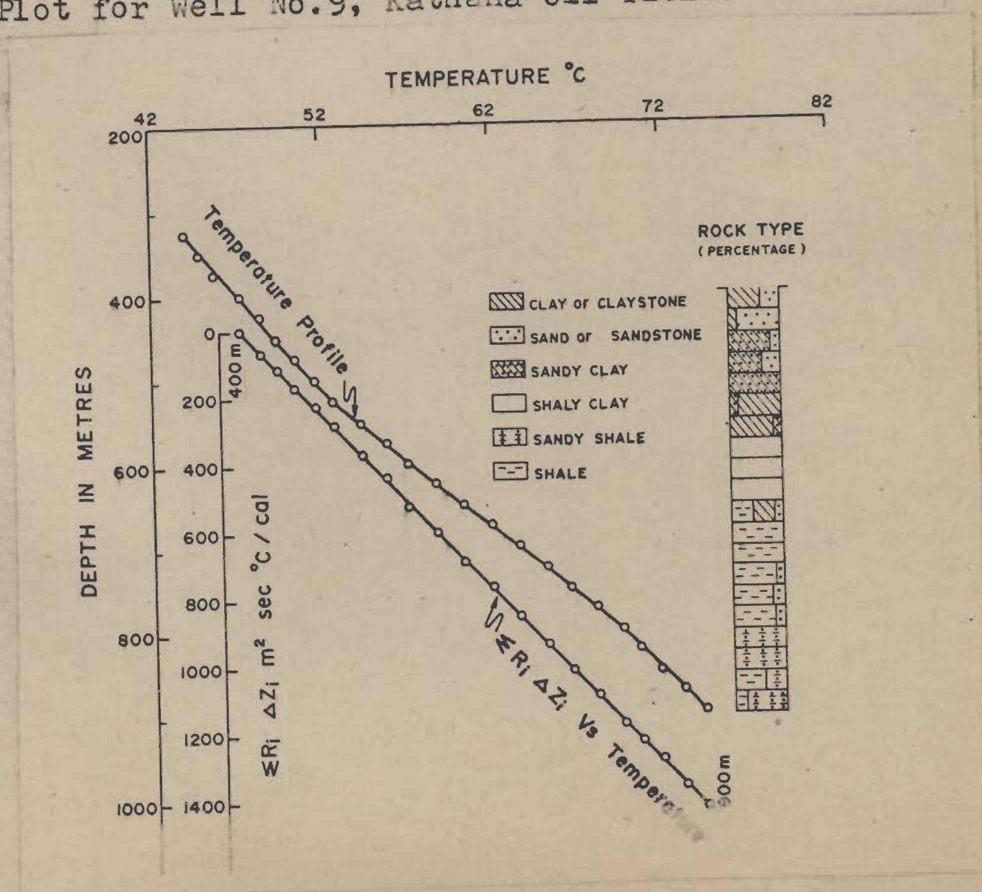


FIGURE III-5

Temperature Profile, Lithology and Bullard Plot
for the Nawagam well No.34

west trend. Temperatures were measured in two Nos. 32 and 43. Well No.32 had a small flow of water from the perforated horizons below the depth of measurements. Howêver, the temperature gradients do not seem to be much affected by the movement of water. The Bullard plot for well No.32 (Figure III-5) is linear from 400-900 m and yields a heat flow value of 1.9 HFU more or less the same as in well No.43 for 600-860 m depth interval (Table III-6).

3.5.4 Sanand oil field

The Sanand structure 16 km WNW of Ahmadabad is a large anticline trending NNW-SSE and having a vertical closure of about 200 m. Temperatures were measured in Well No. S-5 and the Bullard plot yields a heat flow value of 1.8 HFU for the depth interval from 800 to 1000 m in which the temperature-depth curve is quite linear.

3.5.5 The Kalol oil field

Heat flow was determined in two wells, Kalol 27 and 58. Well number 27 is located in the northern part of the Kalol structure, which is an elongated asymmetrical anticline with its longer axis trending in the NW-SE direction (Mathur and Evans, 1964). Kalol 58 is also located in the northern part of the structure and is about 11 km east-southeast of Kalol 27. Temperature prifiles, the percentage of rock types, and the Bullard plots for the two wells are shown in Figure III-6. These were plotted from a depth of about 550 m. The

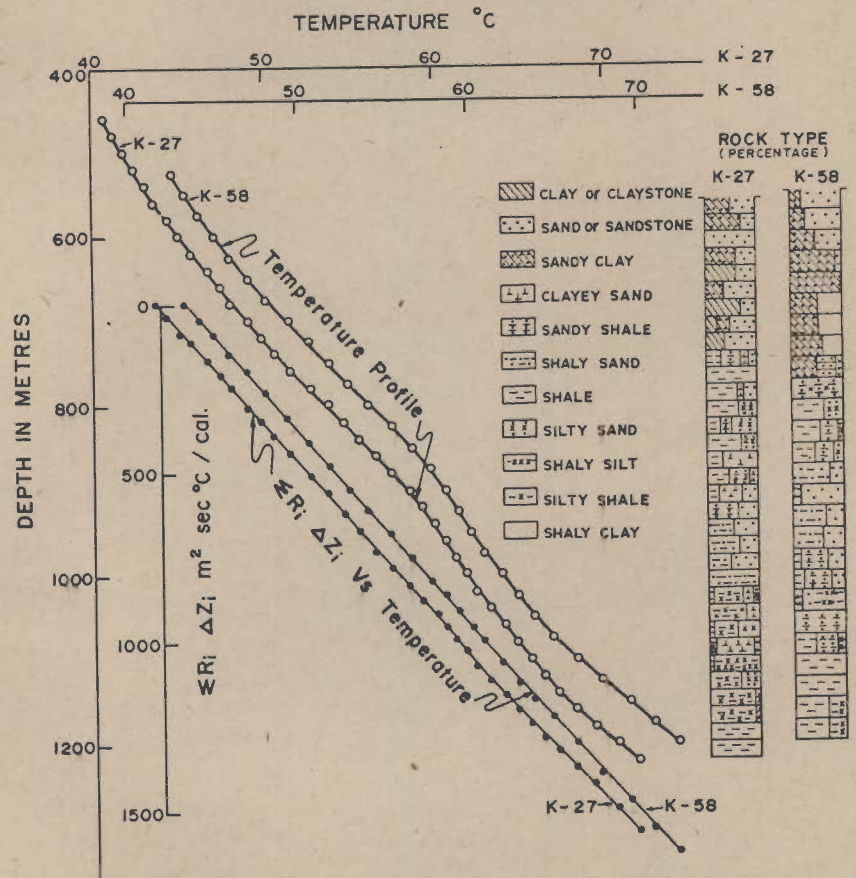


FIGURE III-6

Temperature Profile, Lithology and Temperatures vs $\Sigma Ri \Delta Zi$ Curve (Bullard plot) for Kalol Wells

where ΔZi is the depth interval and Ri is the thermal resistance. The top axis of Temperature is for well No.27 and the bottom axis which is shifted to the right by 2°C is for Kalol well No.58.

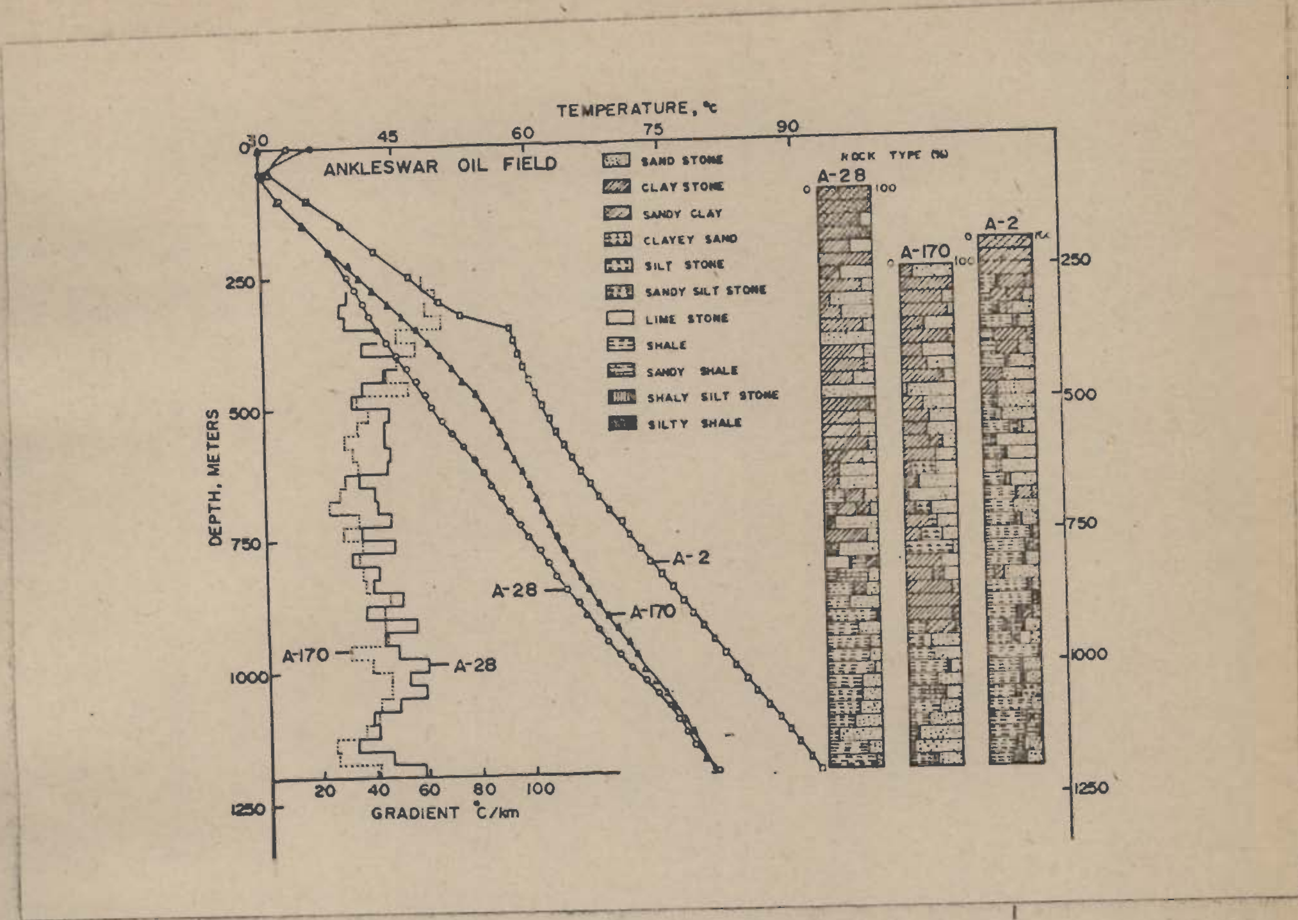


FIGURE III-7

Temperature vs Depth curves, Lithology and Gradients in Ankleswar Wells plots for both the wells show a change in slope near about a depth of 975-1000 m. Therefore, the heat flow were calculated separately for the two different intervals for each well. These are given in Table III-6.

3.5.6 Mehsana structure

Temperature measurements as carried out in well No.6 drilled in Mehsana structure, about 60 km NNW of Ahmadabad, resulted in a heat flow value of 1.9 HFU (Table III-6).

3.5.7 The Ankleswar oil field

The Ankleswar field, 80 km SSW of Baroda, lies on an elongated dome trending ENE-WSW, and is associated with faulting in the trap on its southern flank. The anticline has a gentle northern limb and a steeper southern limb, having a dip that exceeds 20° and even reaches 40°-50° locally.

Temperature measurements were made in three wells A-28, A-170 and A-2 in this field. A temperature disturbance in well number A-2 is likely to be the result of artesian ground water moving up through the well and leaving through horizons near a depth of 350 m. Temperature data of well No. A-170 also indicate a slow upward movement of water through the well. Temperature - depth curves for the three wells are shown in Figure III-7 alongwith interpreted lithologs and temperature gradients each at 25 m intervals. The heat flux

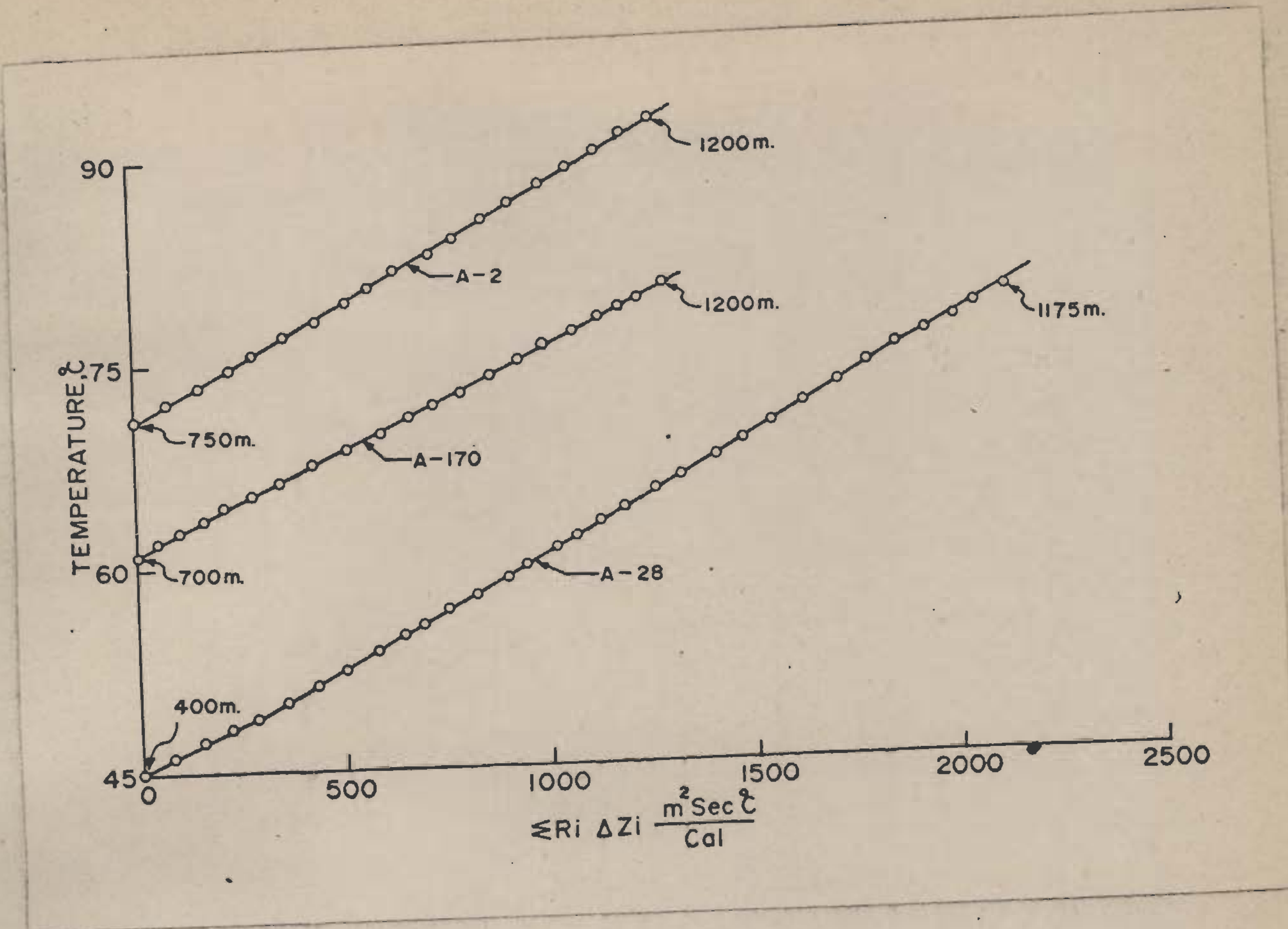


FIGURE III-8

Bullard plots of Ankleswar Wells

has been computed by the Bullard technique for four different depth intervals in these wells and was found to vary from 1.5 to 1.7 HFU (Table III-6). The mean value for the Ankleswar wells is 1.62 HFU. The Bullard plots for the three wells are depicted in Figure III-8.

3.5.8 Broach area

Bottom hole temperature data as collected by the Oil and Natural Gas Commission in one particular well (Broach-6) has been used to calculate the heat flow value for the area. Lowest temperature gradient and heat flow values in the Cambay basin have been obtained for the Broach area (Tables III-2 and III-6). It is worth mentioning that maximum thickness of the sediments in the Cambay basin occurs in the Broach area.

3.6 Discussion of Results

The heat flow values obtained for all the three wells of the Cambay gas field (Table III-5) show considerable variation, the values being particularly lower at above the 850 m level. Mean values of heat flow were calculated for seven depth intervals above the 850 m depth level and for four, below this level. These were find to be respectively 1.76 ± 0.2 and 2.3 ± 0.2 ucal/cm^{2}sec (at 95% confidence level). A similar phenomenon was observed in Kalhana Oil well numbers 3 and 9, where the computed heat flow values increased below the depths of about 650 and 725 m respectively (Table III-6).

These differences arise from: i) instability of the water column upto a depth of 650 - 860 m, or ii) slow underground movement of fluids in the basin at such depths. It may be mentioned here that in the case of Cambay these discrepancies in the heat flows at two levels could not be resolved even when higher thermal conductivity values were assumed for the major rock formations encountered at the upper levels.

Diment (1967) and Gretener (1967) also drew attention to possible instabilities in the water columns in large diameter wells, at depths upto 350 m. Subsequent to the thermal logging of wells and evaluation of heat flow data, an attempt was made to study the temperature variations with time at specific depths in one of the wells. The data although insufficient, owing to lack of facilities for continuous temperature recording at specified depths, show a small order of variation of temperature upto 700-900 m. However it was not possible to predict the presence or otherwise of convection cells in the water column at these depths. Although occurrence of convection cells upto about 800 m can not be ruled out.

Lowering of geothermal gradients and heat flow values in upper levels in account of underground water movements have on the other hand been reported by Bullard (1939) and Bullard and Niblett (1951) and could have played an important role in these regions.

It would appear reasonable to accept the heat flow values obtained at lower intervals as the representative values forthese fields, the final values being given in Table III-1. The error in these values were mainly due to uncertainty in the conductivity values and has been estimated to be ±10%, keeping in view the various experimental uncertanities. The heat flux in the Cambay basin north of the Mahisagar river between Cambay and Mehsana thus appears to be higher than the world average of 1.65 HFU (Horai and Simmons, 1969) and moderate in the southern part of the basin at Ankleswar and Broach. All these values however are considerably higher than the average heat flow values obtained in shield areas.

Since typical heat flow value for stable continental areas is approximately 1.1 HFU (Lee and Uyeda, 1965), the high mean heat flow values of 1.8 to 2.3 ± 0.2 HFU for the various locations in the Cambay basin north of the Mahisagar river call for a sensible elucidation of the heat source which can contribute the additional 1.0 to 1.2 HFU near Cambay. There are no granitic bodies nearby, which could by virtue of their higher radioactivity explain these high values. The Deccan Trap volcanics which underly the sediments are of Eocene age (at least 50 m.y), and are unlikely to be the cause. Heat flow studies in the Deccan Trap region at Alore-Koyna (this chapter, page III-44) and one more location, widely separated from Alore have yielded low values typical of continental stable platform regions. Structural trends indicate that the Aravalli and Pre-Aravalli rocks continue under the basin (S.N.

Sen Gupta of ONGC, Personal discussions). Low heat flow values typical for Archaean shield areas are observed for the area covered by exposed rocks of Aravalli and Pre-Aravalli Super groups (this thesis Chapter IV). According to Watson (1976) almost all the present thickness and volume of the continent was evolved upto 2000-2500 m.y.B.P., and major erosion had taken place only in the first 300-400 m.y. After the elapse of such times the Pre-Cambrian terrains have retained their characteristics. Therefore it can be safely assumed that once the perturbations introduced due to Deccan Trap lava flows die out, similar surface heat flow should be registered both in the Cambay basin and the Pre- Cambrian, Aravalli and Pre-Aravalli terrains, which is not the case.

Unfortunately the source of heat, its spatial distribution in the crust and mantle and its gradual dissipation are not well understood. Vitorello and Pollack (1980) suggest that 30% of the heat flow in Cenozoic Tectonic zone is a residual heat from a transient thermal perturbation associated with tectogensis. During tectogensis a portion of crust and upper mantle become thermally disturbed by the process culminating in metamorphisim, magmatism, deformation and uplift. The heat anomalies of geosynclinal areas such as Cambay basin have evidently resulted from tectonic activization. They are caused by deep seated processes and are sometime revealed in the upper crust by basic and acid volcanism, the carriers of heat energy and the products of deep seated melting and differentiation. In geosyncline areas the heat field is not homogeneous and areas of high and low heat flow are distinguishable. In some cases high heat flow is found in areas of strong magmatic activity, positive gravity anomalies, lower seismic velocities and higher electrical conductivity of the mantle. Weak magmatism and negative gravity anomalies occur with the low heat flow.

A composite Bouguer anomaly map of the Gujarat area has been presented by Sen Gupta (1967) (Figure III-9), which shows an interesting gravity high along the axis of the basin, running almost through the Central part of the Cambay basin in the area north of the Mahisagar river from Cambay to near Patan (Figure III-9). The prevalence of negative anomalies in the area towards south of Mahisagar river is also quite evident. The Bouguer anomaly has a maximum of about +37 mgal near Cambay, +17 mgal near Sanand and decreases to about +8 mgal further northwards near Kalol, just as the heat flow anomaly decreases. The gravity high cannot be explained by any structure within the sediments which are of low density, 2.3 to 2.4 gm/cm^3. Kailasam and Qureshy (1964) have explained this high near Cambay as being due to greater thickness of Deccan traps in the central part of basin as compared to its margins. Assuming a density contrast of 0.3 gm/cm^3 between the Deccan traps and the adjoining crustal rocks, they have obtained a thickness of Deccan traps of 4 to 4.5 km in the central part of the basin. Although this interpretation could explain the gravity high in the basin, it cannot explain the high heat flow. The combined thermal and gravity highs can be caused

FIGURE III-9

Bouguer Anomaly Map of Cambay Basin after Sen Gupta (1967) due to shallow crustal intrusion, and or a relatively crustal thinning beneath the basin. Deep seismic soundings have not shown appreciable crustal thinning in the main parts of the Cambay basin (Kaila personal communication). As has been mentioned that the gravity high is confined only to the axis. Therefore the likely source for the high heat flow in the basin could arise from an intrusion in the earth's crust.

In order to estimate the parameters of a causative intrusive body, a model of an infinite rectangular dyke of cross section 50 x 50 sq.km and original temperature 1200°C was assumed. Heat flow values were calculated using the method suggested by Simmons (1967) for depths of burial at 0, 10, 20 and 30 km. Figure III-10 shows the relationship between the heat flow contribution and age of an intrusive body of the above size. It appears that a heat flow anomaly of about 1.0 to 1.2 HFU could be explained as being caused by an intrusion of about 10 m.y. age at a depth of about 10 km. The igneous intrusion which must be of late Miocene or Pliocene age might also have been responsible for faults and other tectonic features found in the Cambay basin.

Further support regarding the possible intrusion in the earth's crust at a shallow depth underneath the Cambay field has been obtained from occurrence of steam in some wells. Sen Gupta (1 967) reported encountering high pressure steam in two wells, one each in Cambay and Kathana, at depths of about 1500 to 1750 m. All evidence points out to the source of high temperature steam being within the Deccan Traps. Steam

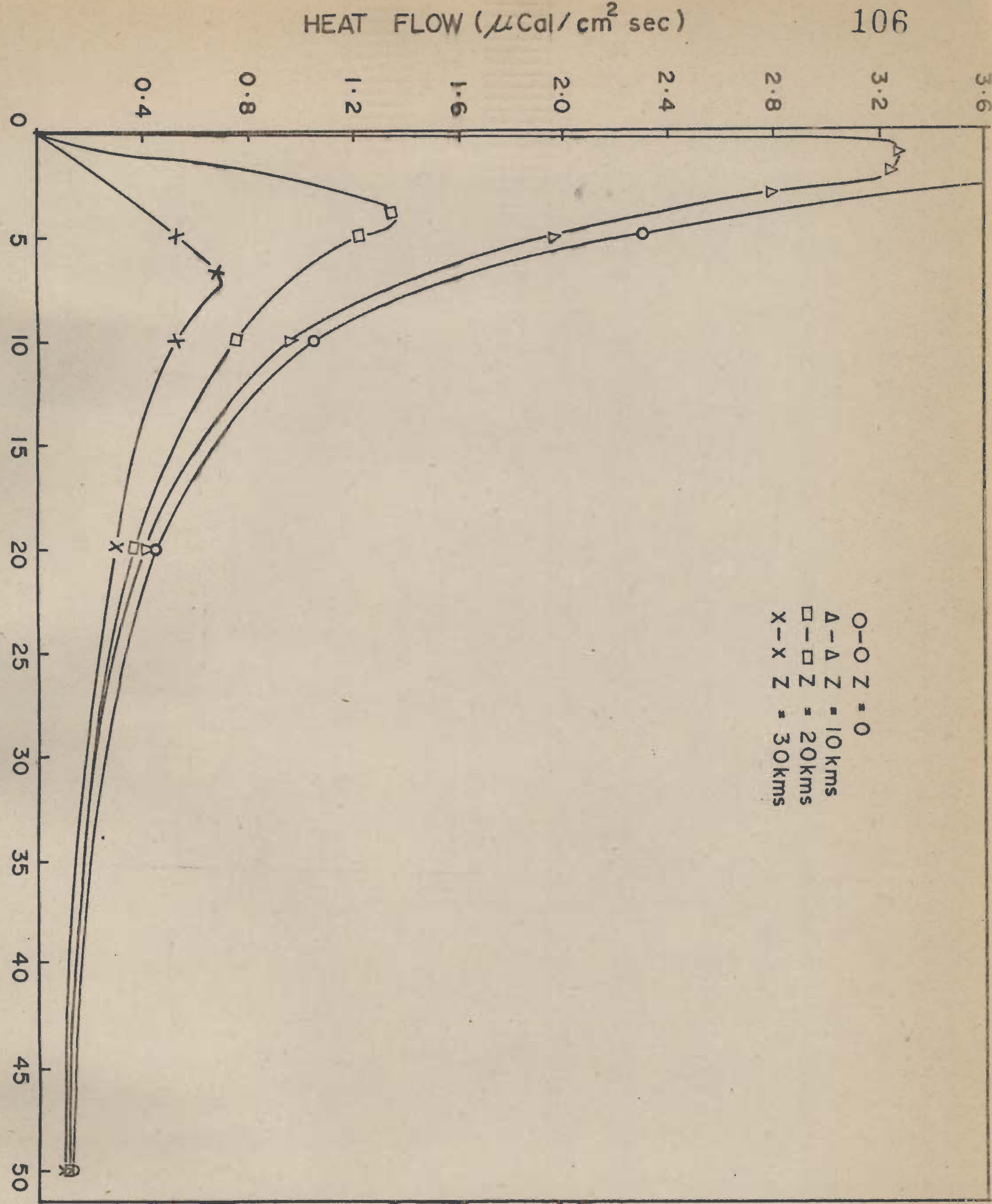


FIG. III-10 Relationship between heat flow and age of an intrusive body of 50 x 50 Km, dimension at a depth of 0,10,20,30, Km below the surface.

has not been encountered in other wells drilled to the same depth in this same area. It appears that the high temperature steam must have existed as a water phase within the traps and found its way up along faults extending through the sediments into the traps below. The high temperature of water inside the traps might have been maintained by the heat of the underlying intrusion and the two wells drilled near Cambay encountering these deep seated faults provided an escape passage for the steam.

The magnitude of heat flow anomaly decreases as we proceed from Cambay towards Kalol-Mehsana. It appears probable that the intrusive body, suggested above has varying altitudes within the crust underneath the Cambay field and extends from Cambay to Mehsana. As mentioned earlier the gravity data of the Cambay basin lends support to such an assumption. The consistent trend of the two fields appears to suggest that the intrusive body narrows towards the north.

Assuming the common origin of these two fields a model of the E- W crustal section near Cambay has been constructed with density contrasts as given in Figure III-11. For the most part the Deccan traps have been assumed to overlie the Archaeans, as appears from geological evidence. Crustal rocks are assumed to consist of a lighter density material near the surface and higher density on account of overburden at greater depths. A gravity profile AA_1 (Figure No. III-9) in ENE-WSW direction passing near by Cambay has been thus interpreted. The gravity profile is asymmetrical and

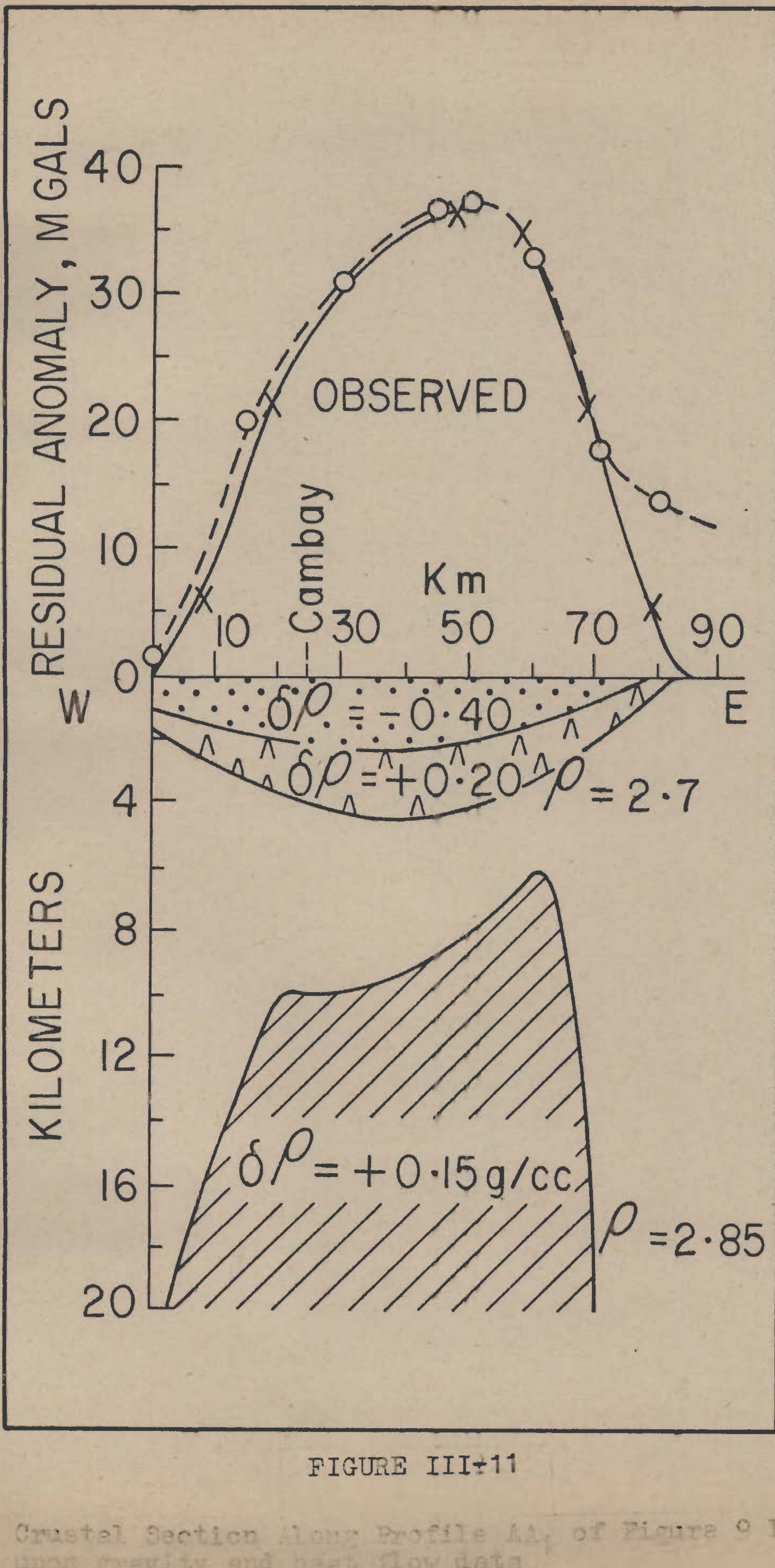


FIGURE III-11

Crustal Section Along Profile AA, of Figure 9 Based upon gravity and heat flow data shows a residual high of the order of 37 mgal, about 25 km ENE of Cambay.

Assuming the gravity high to be only due to the intrusive body in the crust with a density contrast of +0.15 gm/cc, the approximate dimensions of the body were obtained using Skeel's (1963) two dimensional method. The depth to the top surface of the body (D_1) was thus found to be 11 km and that to the bottom (D_2) 20 km. Its width turned out to be 44 km. Assuming a slightly lesser density contrast of 0.1 gm/cc, the dimensions obtained were $D_1 = 10.3$ km, $D_2 = 25.8$ km and $W = 43.8$ km. A lower value of density contrast therefore results in increasing the interpreted vertical dimensions of the body but not so much its other dimensions. Starting with approximate dimensions of the body as obtained by Skeel's method and the sediment configuration as reported in literature, Hubert's (1948) line integral method was applied to obtain the shape of the intrusive body and that of the underlying ~~x~~ traps. An accurate estimate of the thickness of the traps in the basin is not availabe but it is likely to be 1 -2 km as is indicated by wells drilled by ONGC. For purposes of computation, the thickness of the traps under the sediments in the central part of the basin was taken to be about 2.2 km. The final shape of the intrusive body, the traps, and the sediments which can explain the observed gravity curve most satisfactorily are shown in Figure III-11. The discrepancy between the observed and the calculated curves near the eastern margins of the basin appears to be due to the presence of a secondary structure along the eastern margins. Considering the uncertainties in our knowledge of the density contrasts and in the consequent interpretation of gravity data, the crustal model presented here, based on existing data, appears to be the most likely one.

A prominent gravity anomaly of magnitude 8 mgal covering a wide area of about 800 sq.km centred near Borsad about 25 km northeast of Cambay, has been attributed by Negi (1952) to a probable change in density within the earth's crust. A strong magnetic high (500 gammas) slightly displaced to the southeast, has also been reported in this area which is probably of deep-seated origin (Ramachandra Rao, 1958). These observations further support the probable presence of an intrusive body.

Another conspicuous feature in the Bouguer anomaly map, Sengupta (1967), is the absence of a gravity high in the southern part of the Cambay basin. The heat flow values which could be obtained in this part of the basin at Broach and Ankleswar are also relatively lower. However, for a better understanding, more heat flow data is necessary in the southern part of the basin.

As mentioned earlier, reversals of block tilting have been a conspicuous feature of the basin throughout its Cenozoic history. The Cambay-Kathana region has acted as a central higher zone between two areas of active subsidence both towards the north and south. The highest heat flux is observed in this central zone.

3.7 Correlation With Basement Topography

The decreasing trend in heat flow values from Cambay to Kalol also bears a correlation with the basement topography. A map of the trap surface in the Cambay basin has been prepared by Avasthi et al. (1969) vide Figure No. III-2. The depths to the trap surface for different fields north of the Mahisagar river are given in Table III-7. The table indicates a decrease of heat flow with increase in the depth of the basement complex.

? 71

TABLE III-7

Basement depths and average heat flux at different fields in the northern part of the Cambay Basin

Locality	Basement depth (m)	Average heat flow u cal./cm^2.sec
Cambay gas field	2,200	2.3
Kathana Oil field	2,400	2.2
Nawagam Oil field	2,700	1.9
Sanand	2800 - 3000	1.9
Kalol Oil field	3000 - 3200	1.85

3.8 Conclusions

The above perspective of basement topography, tectonics and geophysical anomalies delineate certain revealing features which are briefly summarised below:

i) The Cambay basin area north of the Mahisagar river from Cambay to Sanand is a region of high heat flow. The association of high heat flow values with gravity 'high' in this part of the basin indicates the likelihood of a buried igneous intrusion of Pliocene to Miocene age in the crust underneath the basin, which probably extends from Cambay towards Kalol and Patan.

ii) The Cambay-Kathana region which apparently acted as a central hinge zone separating the two areas of active subsidence in the north and south during the Eocene Epoch, is found to be one of highest heat flux in the Cambay Basin. The maximum gravity and magnetic anomalies found in the region also appear to be associated with this zone.

iii) There seems to be a clear correlation between the heat flux and the basement relief which may be associated with oil and gas structures. Geothermal gradients and heat flow are higher where the depth of the basement complex is less and vice versa.

3.9 Heat Flow at Alore-Koyna, Deccan Traps

Alore and Koyna are located in the Western part of the Maharashtra state within one of the most important geological units of Peninsular India - the Deccan Traps. Hydroelectric power is generated in the area by impounding water from the river Koyna at a high altitude in a lake named Shivaji Sagaar. Two geomorphic patterns are delineated here: i) the Sahyadri mountains, which form a system of ridges and ii) a gently sloping narrow coastal country from the West of the continental divide to the Konkan coast. The Sahyadri Mountains act as a water shed between the Konkan and the Eastern mainland area. Koyna and Alore are respectively located towards the Eastern and Western side of the water shed (Figure III-1 2).

Although the area lies in the largely aseismic peninsular shield, it began to show mild seismic activity soon after the impounding of the Shivaji Sagar Lake in 1962. Microearthquakes progressively increased thereafter, both in frequency and magnitude, over the next five years; and after the power production reached its peak in the year 1967, an earthquake of the magnitude 5.8 on the Richter's scale occurred in the area on September 13, 1967. This was followed by a major shock of magnitude 7.0 on December 11, 1967, and the increased level of seismicity has continued upto this day. The epicentres of both these earthquakes were located in the neighbourhood of Koyna nagar. The occurrence of an earthquake

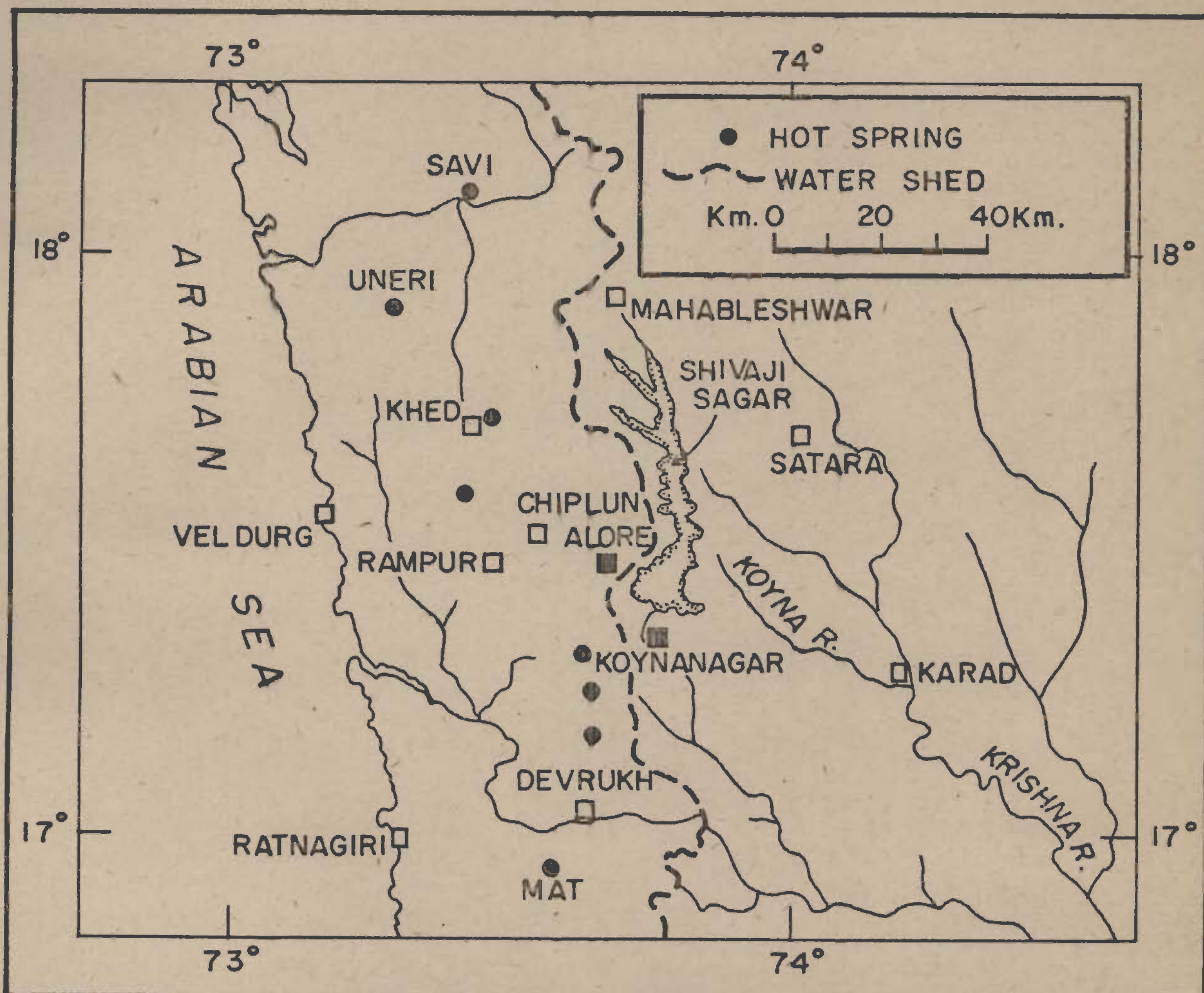


FIGURE III-12

Location map of Alore-Koyna Area

of magnitude 7.0 in the area became an important land mark and triggered a lot of scientific investigations. Two boreholes were drilled near Alore, by the Koyna Hydroelectric Project Authorities for estimating the thickness of the traps in the region which were subsequently used for temperature measurements reported here.

3.9.1 Geology and tectonics of the Alore-Koyna region

The Deccan trap formations, over which Koyna and Alore are located, constitute a series of basic lava flows of Eocene-Paleocene age. The close of Mesozoic era in India was marked by the outpouring of enormous lava flows, which spread over vast areas of Western, Central and Southern (Deccan) India, as nearly horizontal sheets that have resulted in their terraced form. These flows cover an area of about 600,000 sq.km attaining a maximum thickness of nearly 10 km(?) in the region of the Gulf of Cambay and gradually thinning out eastwards to just a few metres.

Thicknesses of the individual flows vary from a metre to over 50 metres. Some of these have been traced over long distances exceeding 100 km. Three types of basalts have been recognised viz., Tholeitic, high alumina basalts and alkali-olivine basalts. Numerous dolerite dykes which are not evently distributed are also known to occur in the Deccan Trap. Both, post-Trap hypabyssal intrusives and feeder dykes for basalt flows, have been observed. It is believed that the Deccan Trap lava flows erupted from a system of a fissures in the Earth's crust (West, 1958). However no clear cut evidence about these fissures has yet been observed. Biswas and Deshpande (1970) opine that the Deccan Trap lava might have also erupted from the shield volcanoes situated along the west coast and in the Cambay and Narmada-Tapti rift zones. Wellman and McElhinny (1970) based on the K/Ar ages of some Deccan Trap samples, suggest that they are of Paleocene age and the actual span of volcanic activity is of the order of 5 million years, with an average age of around 60 m.y. Several evidences of post-Trappen faulting, rejuvenation of the plateau, accumulation of tectonic stress and of recent faulting in the coastal strip have also been observed (Sahasrabudhe and Deshmukh, 1970; Gupte, 1970; Pawde and Kumar, 1976; Das and Ray, 1971).

A belt of hot springs are known to occur on the Konkan coast which runs for over 450 km as a narrow belt towards the West of the continental divide. From photogeological studies, Das and Ray of the Geological Survey of India conclude that the hot springs of this belt follow a definite tectonic alignment and are related to fracture systems in the region.

According to S.C. Sharma of the Geological Survey of India who carried out detailed geological mapping and field surveys in the Alore-Koyna areas, the region has perhaps not suffered any severe tectonic movements in the recent past and whilst there exist a large number of N-S trending fractures west of the continental divide, they are comparatively fewer in number on its east. He also noticed the presence of numerous shear zones in the excavations made in connection with the Koyna Hydroelectric project.

Temperature measurements were made in two boreholes drilled at Alore. These encountered dense grey and Amygdaloidal varieties of basalts. Intertrappeans and pyroclastics were not encountered as expected, indicating thereby that the volcanic activity might have continued uninterrupted and without violent outbursts. The proven depth of the basalts is 214 m below the m.s.l. (as borehole KFE 2 was drilled upto this depth). No marine sediments were found in association with the pillow structures to suggest that these lavas had been laid under the sea, and it appears reasonable to assume that the area was downthrown by more than 214 m (S.C. Sharma, personal communication).

3.9.2 Temperature measurements

Temperatures were measured to within ±0.02°C in two boreholes KFE-1 and KFE-2, at various depths at 5 m intervals. Temperatures in KFE-1 were measured a few years after the drilling had been completed while in KFE-2 these were made only two weeks after the drilling. Gradients measured at 5 m intervals and the temperature - depth curve and litholog for the two boreholes are shown in Figure III-13.

3.9.3 Measurements of the coefficient of thermal conductivity

Conductivity coefficients of 24 core samples collected

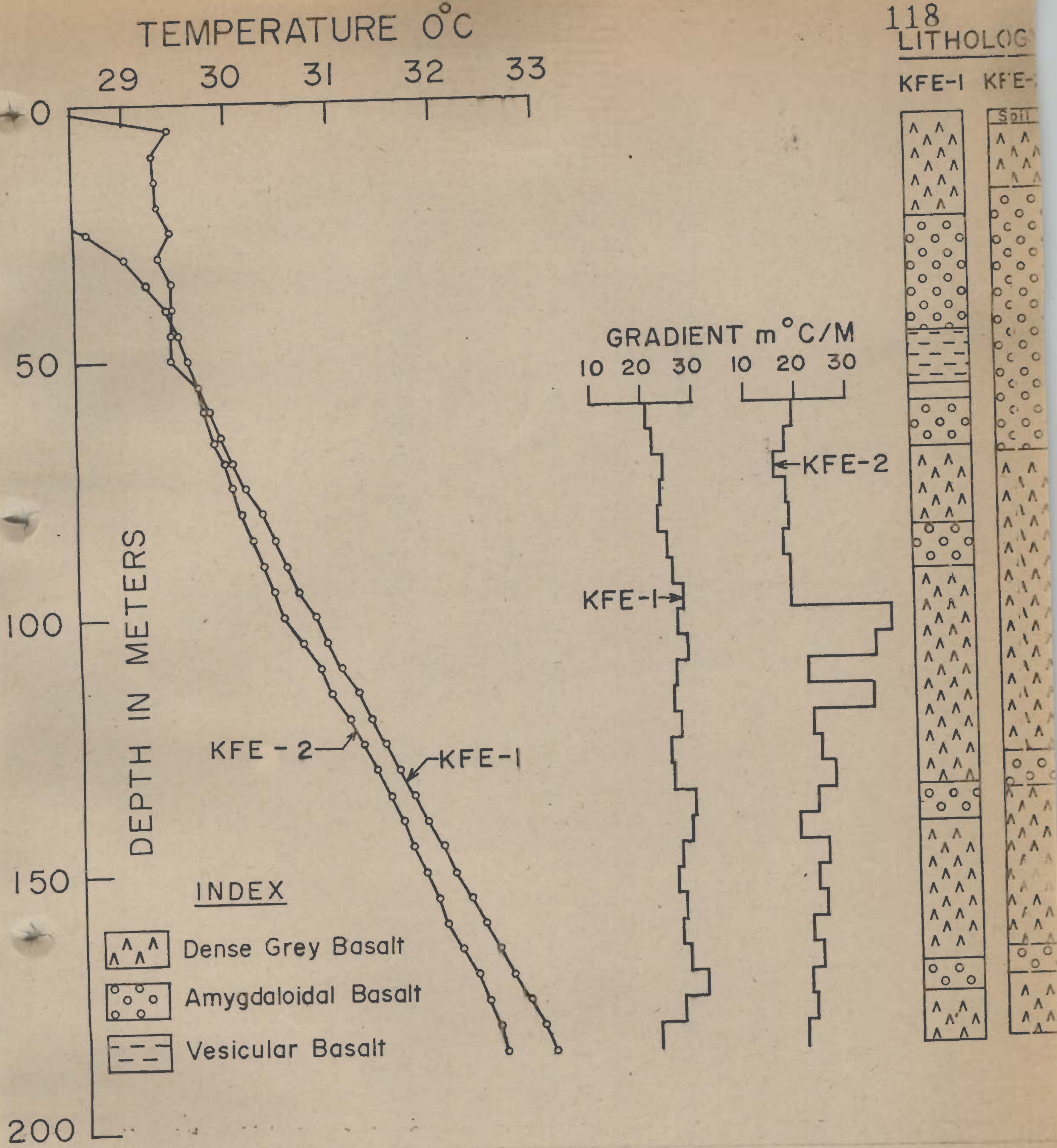


FIG. III-13 Temperature Vs Depth Curves, Lithology and Gradients in Koyna Wells.

from the two boreholes were measured using the modified Birch type apparatus and the values are given in Table III-8. During measurements the samples were maintained at a mean temperature of 37.5°C and under a pressure of about 100 kg/cm^2.

Upon a microscopic study, the core samples were classified into grey basalts and Amygdaloidal basalts which were partly vesicular and included secondary mineral veins and grains. Some basalts show predominent inclusions of Calcite as veins and grains. The densities of the samples varied from 2.29 to 3.04 gm/cm^3 and the conductivity values from 2.2 to 4.6 mcal./cm.Sec.°C. Thermal conductivity of dense grey basalts was found to be higher than those of the amygdaloidal variety.

3.94 Evaluation of Heat Flow

Heat flow values have been evaluated for two depth intervals: 75 to 130 m and 130 to 180 m in borehole IFB-1. In the first interval the gradient was found to be 25.98 ± 0.38°C/km which yielded a heat flow value equal to 0.96 HFU, when combined with the mean conductivity of all the core samples. The gradient in the second interval was 27.2°C/km yielding the corresponding heat flow of 1.01 HFU. Similarly the gradient and heat flow values for the two depth intervals 95 to 140 m and 145 to 170 m in borehole KFB-2 were found to be 27.74°C/km, 23.27°C/km and 1.03 HFU and 0.86 HFU respectively. However, at lower depth intervals in borehole KFB-2,

TABLE III-8

Thermal Conductivity Data for Alore-Koyna Boreholes

Rock Type: Dense grey basalts

S. No.	Bore hole No.	Mean depth (m)	Thermal conductivity (mcal/cm^2. Sec °C)
1.	KPB-1	68.5	3.83
2.	-do-	129.0	4.49
3.	-do-	136.5	3.86
4.	-do-	142.6	3.58
5.	-do-	148.7	3.78
6.	-do-	162.4	3.80
7.	-do-	163.0	4.62
8.	-do-	168.5	4.08
9.	-do-	171.5	4.39
10.	-do-	172.0	3.74
11.	-do-	177.6	4.02
12.	-do-	182.8	4.52
13.	KPB-2	85.3	3.50
14.	-do-	95.0	4.47
15.	-do-	142.5	4.18
16.	-do-	161.3	4.45
17.	-do-	179.0	3.81
	Mean :		4.07

Rock Type: Dense grey basalt with few amygdules

S. No.	Bore hole No.	Mean depth (m)	Thermal conductivity (mcal/cm^2. Sec °C)
1.	KPB-1	171.5	3.47
2.	KPB-2	73.0	3.67
3.	-do-	113.0	3.32
	Mean :		3.49

Rock Type: Amygdoloidal vesicular basalts altered

S. No.	Bore hole No.	Mean depth (m)	Thermal conductivity (mcal/cm^2. Sec °C)
1.	KPB-1	63.9	2.84
2.	KPB-2	129.8	3.21
3.	-do-	172.0	2.97
	Mean:		3.01

Rock Type: Coarse grained basalts with vesicular filled with calcite

S. No.	Bore hole No.	Mean depth (m)	Thermal conductivity (mcal/cm^2. Sec °C)
1.	KPB-1	98.0	2.59
2.	KPB-2	136.4	2.22
	Mean :		2.40

dense grey basalt and fresh hard amygdoloidal basalts were mostly encountered both possessing higher conductivity values. The mean conductivity for these varieties of basalts (4.07). When combined with the gradient of 23.27°C/km as observed in the lower depth interval of KPB-2 yields a heat flow value of 0.95 HFU.

It was difficult to work-out precisely the thicknesses of various types of basalts encountered in the two boreholes. Observed variations in heat flow appear to arise mainly from random variations in conductivity. The mean heat flow obtained from the four depth intervals is 1.0 HFU.

3.9.5 Discussion

The heat flow value of 1.0 HFU for Alore is typical of continental platform regions. Since the Deccan Traps have blanked the pre-existing topography over a large area and lie juxtaposed with the Pre-Cambrian Dharwars, granites, gneisses, Kaladgi and Bhima Series sediments and Gondwana sedimentary formations, the stratigraphy of the underlying formations was a matter of wide speculation. It is however generally believed that the Deccan Traps overlie the Pre-Cambrian formations most of the way. The Deccan traps have a mean age of about 60 m.y. and have lost practically all their volcanic heat. Estimates of the inferred thickness of basalts in the Alore region vary from 250 to 450 m. Also, since basalts posses low concentrations of radioactive heat generating elements, most of the Alore heat flow represents characteristics of a Pre-Cambrian crust. Observed heat flow values are respectively 0.7 and 0.6 HFU in Dharwars south of Alore (Chapter IV); 1.49 HFU at Damua and 1.18 at Mohapani in the Gondwanas towards the northeast.

Although the heat flow value obtained from Alore constitutes a single sample, it appears to be reasonably free from any disturbances. The value is similar to model values from other stable continental areas (Lee and Uyeda, 1965). If it is assumed to be a representative value for the area, it indicates that the shallow seismic activity of the region, which may have been caused by reactivation of some local shallow faults and readjustment after the impounding of Shivajisagar lake, does not appear to be of much tectonic significance.

CHAPTER IV A

HEAT FLOW IN ARAVALLI MOUNTAIN BELT AND AT KALYADI (DHARWARS)

4A.1 Introduction

The Aravalli mountain system constitutes an important Pre-Cambrian geological unit of the Indian landmass. It occupies the north-western part of the Indian shield has an average width of about 80 km and extends over a distance of nearly 600 km from Delhi to Gujarat across Rajasthan in a south-westerly direction (Figure IV-1). Near Gujarat it assumes a south-easterly trend and disappears under the Deccan Traps. It may in fact be extending from underneath the Himalyan architecture in the north to the Laccadives in the south (Krishnan, 1953).

The Aravalli range is the richest base metal province in India encompassing a large number of deposits and an unusual concentration of copper and lead-zinc mineralisation. Considerable exploration and drilling activity has consequently been carried out there. Heat flow measurements in three deep boreholes at Madankudan in the Khetri area yielded surprisingly high surface values of 1.76 HFU (Gupta et al. 1967), thereby stimulating interest to study as many heat flow values as possible over its entire exposed length. Temperature measurements were accordingly made in a large number of diamond-cored holes of varying depths. Numerous cores were sampled for

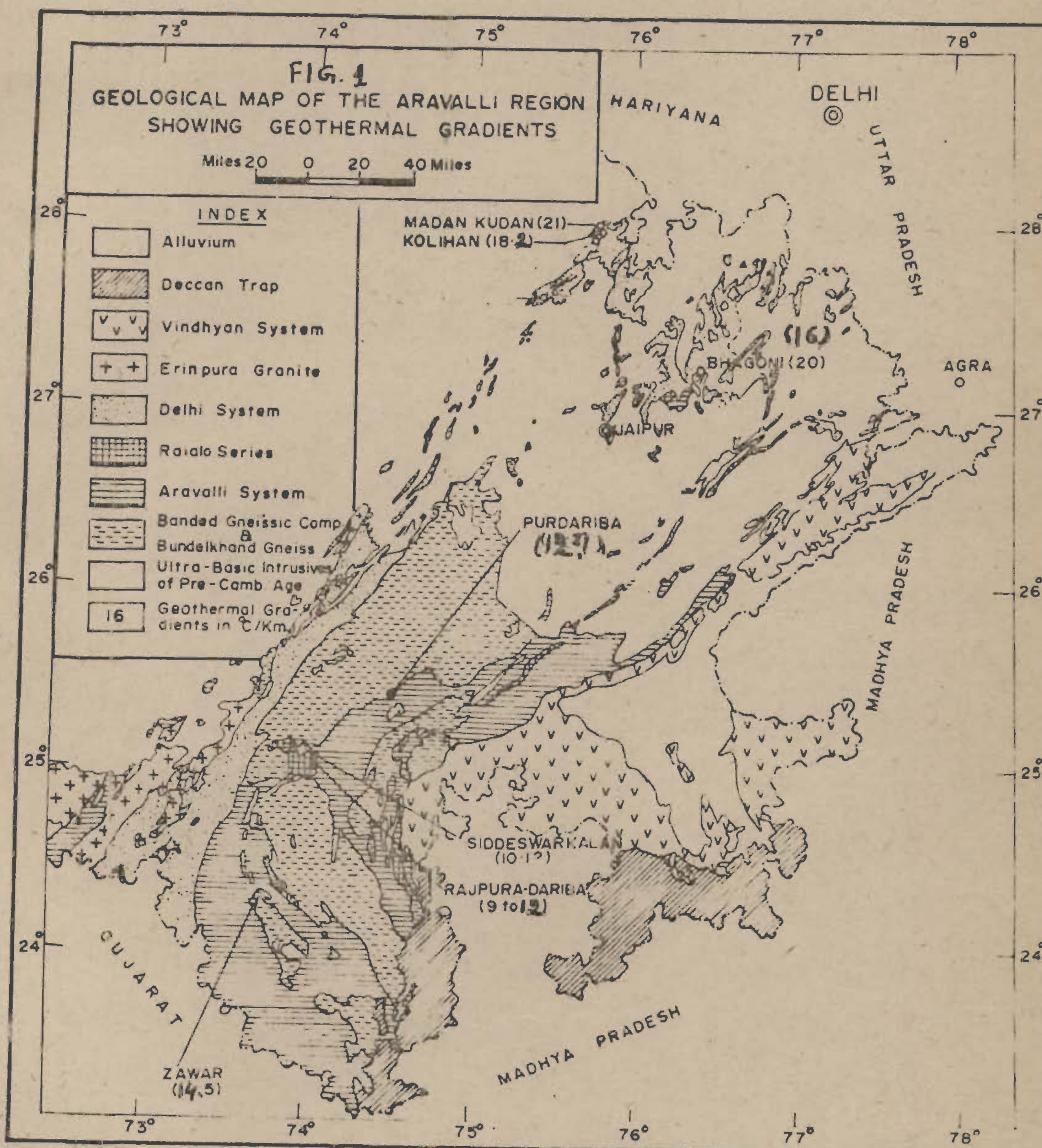


FIG. 1
GEOLOGICAL MAP OF THE ARAVALLI REGION SHOWING GEOTHERMAL GRADIENTS

conductivity measurements in the laboratory to evaluate heat flow values.

Results obtained in the Madan-Kudan, Kolihan, Kala-pahr, Bhagoni, (Delhi-Super Group), Zawar (Aravalli Super Group), Rajpura-Dariba, and Pur-Dariba (Pre-Aravalli Super Group) areas of the Aravalli mountain system (Figure IV-1) as well as heat flow value obtained at Kalyadi, in a Dharwar schist belt are presented and discussed in this chapter.

4A.2 Geology of The Aravalli Belt

The Aravalli belt is composed of three Pre-cambrian Super Groups: the Pre-Aravallis, Aravallis and the Delhis.

The Pre-Aravallis constitute the oldest rock complex of the belt and include a wide variety of crystalline rocks represented by schists, phyllites, slates, quartzites, marbles and gneisses with associated acid and basic intrusive bodies. The Bereach granites and gneisses, the oldest member (2580 m.y) of the Pre-Aravallis, are exposed in the central part of Rajasthan and resemble, over a large part of the area, the post-Dharwarian granite (Closepet type) of South India. The Banded Gneissic Complex of eastern Rajasthan, overlying the Bereach granites, is represented by migmatites and composite gneisses that undoubtedly include original sediments, now represented by relict bands, and lenses of biotite and chlorite schists, (Banerje and Mitra, 1977).

Rocks of the Aravallis super group which are comparatively less metamorphosed than the underlying Pre-Aravallis and the overlying Delhis, comprise a thick folded sequence of argillaceous sediments represented by quartzites, arkose and conglomerates at the base and mica schists, quartzites, phyllites, cherty limestone, dolomite and composite gneisses in the upper part (Muktinath, 1967). The basal beds are found to rest on the Banded Gneissic Complex with an erosional unconformity. The thick series of ferruginous and argillaceous dolomite and limestone beds exposed between Kankroli, Nathdwara, Udaipur, Zawar and Dungarpur areas, which were earlier considered by Heron (1925) as constituting the Raialo, are now grouped under the Aravallis as they represent a continuous sequence.

The overlying Delhi super group of rocks are well developed along the main axis of folding of the Aravalli range an extend from near Delhi in the northeast to Gujarat in the southwest (Figure IV-1). Its lowermost group, consisting mostly of limestones, with conglomerates and quartzite at the base has been designated as the Raialo Group by Banerjee and Mitra, (1977), the name 'Raialo' being retained only in the type area where it rests on the Pre-Aravalli Banded Gneissic Complex and is overlain by the Alwar Group, both the junctions being unconformable. The Alwar group is composed of arkoses and grits e.g., phyllites, schists; Arkose quartzites, described as Alwar series, at the base, followed by the upper most series named the Ajabgarh series or group. The Ajabgarh group is

composed of a

composed of a thick pile of metasediments represented by phyllites, biotite schists, calc-gneisses.

All these formations of three super groups and sub-groups are associated with a number of acidic, basic and ultrabasic intrusives of different Pre-Cambrian ages and rich base metal deposits. The eastern part of the Aravalli embrases the major part of the deposits. The entire stretch of the metallogenetic belt has a remarkably NNE-SSW linear trend i.e., parallel to the regional strike. It is considered to be a structurally weak zone and constitutes a major tectonic lineament that might have originated due to long continued deformative movements. Other minor linear fabrics, in the form of linear faults, fracture and shear planes, are also associated with the major one and essentially run parallel to it.

The metallogenic-minerogenetic epochs of the Aravalli region have been closely associated with its three main orogenies viz., Pre-Aravalli, Aravalli and Delhi of the region. The major mineral belts such as Khetri and Alwar copper belts, Zawar lead-zinc belt, and Bhilwara lead-zinc belt, are respectively related to the Pre-Aravalli, Aravalli and Delhi orogenies (Muktinath, 1967).

4A.3 Evolution of The Aravalli Belt

Modern views regarding the initial growth pattern of continental masses envisage nucleation, accretion, and merger into protocontinents (Goodwin, 1960; 71; Anhaeusser, 1973; Glikson, 1971). From the discovery of charnockites and associated granulits in the 'Banded Gneissic Complex' in Bandanwara area (Ajmer) by Pandaya (1970) and Gyani (1970), existence of a nucleus of the Katarchean continent of western and central north India has been inferred in the Bundelkhand and Ajmer-Mewar region of the Aravallis by Raja Rao (1971) and Valdiya (1973).

The oldest Katarchean rocks (3,000 m.y) of the Aravallis are represented by enclaves of old metamorphics present within the massif of the Bundelkhund Granites, and by the gneisses, granulites and charnockites of Central Rajasthan that have been extensively migmatised or otherwise reconstituted into the Banded Gneissic Complex. Crawford (1970, p.105) has shown that the Bereach as well as the Bundelkhand Granites have the same age (2555 ± 55 m.y). The Bundelkhand orogeny, regional metamorphism, and granitization probably concluded about 2450 m.y. ago. This is supported by the fact that the overlying Bijawar lavas have a total rock Rb-Sr age of between 2460 and 2510 m.y. (Crawford, 1969, p.381). A Pb-isochron age of 3500 m.y. has been reported from the Aravalli schists 8 km east of Udaipur, which according to Pitchamuthu (1971) probably indicates the age of some component of the basement complex.

All the lithological formations of the Aravalli range mentioned in section 4.1, are constituted of folded metasediments of ancient seas. The range is the result of crustal movements in the Purana era, although it now appears as a minor feature, it was once very lofty. Ever since the Precambrian times, it has been subject to continued denudation, except for some upheaval during the Mesozoic era, of as Large an order as 24 miles(?) (Gulatee, 1952). On geological evidence, the Aravalli range appear to have been peneplaned in Pre-Crateceous times.

Detailed structural studies of the Aravalli region notably by Ghosh and Naha (1962), Naha et al. (1968), Naha and Majumder (1971), Sen (1970), Banerjee and Mitra (1977) etc., revealed that four successive cycles of non-affine deformative movements, followed by later minor orogenic impulses, spread over a long span of time have occurred. The Delhis are characterised by at least two sets of major folding movements giving rise to NNE-SSW and WNW-ESE structural trends, while the Aravallis are associated with two completely seperate structural trends, i.e., E-W and NW-SE, which indicates that the tectonicity varied in space and also in time during the Delhi and Aravalli times, the fold-pattern in the Pre-Aravallis being the same as that in the Aravallis, thereby confirming an identical tectonic hierarchy prevailing during these two periods. The main metamorphic episodes in the Pre-Aravallis, Aravallis and Delhis, which were, sensu stricto regional in nature, were coeval with the earlier phases of the major folding movements and were followed by dislocation and diaphthoresis in zones of shearing, thereby causing retrograde metamorphism in limited locales. The grade of regional metamorphism, in general, varies between lower greenschist facies to Upper amphibolite facies. The presence of granulites, at places, however, suggests higher temperature and pressure conditions prevailing at least locally, (Banjerjee and Mitra, 1977).

Pegmatites occur during all the three orogenies and the difficulty in associating these nth specific orogenies have led to a certain amount of uncertainty in the interpretation of geochronological data. However, on the basis of the pioneering work of Holmes (1949, 1950, 1955), the K-Ar ages of Vinogradóv and Tugarinov (1964) and the recent ages determined by Crawford (1969, pp.380; 1970, pp.108) the following sequential events have been recognised.

A whole-rock Rb-Sr isochron age of the basement complex near Chitor is 2600 m.y. This probably marks the closing of the orogeny, metamorphism, and granitisation of the Pre-Aravalli basement in Rajasthan. According to Crawford the base of the Aravallis can be no older than 2500 to 2590 m.y. According to Pichamuthu (1971) it need not be quite so old if there was a large time gap between the crystallisation of the Bundelkhand and Bereach granites and the deposition of the Aravalli sediments. The base of the Aravallies is taken down to 2100 m.y. (Muktinath, 1967). Aravallis are intruded by granites with a minimum age of 1900 m.y. The age of the top of the Aravallis must be of this order or less. According to Sarkar (1968) the Aravalli orogeny and metamorphism closed between 950 and 1500 m.y.

The Delhi super group is intruded by the Bairath Granite dated at about 1660 m.y. thus suggesting that the age of the Alwar series of the super group and probably of the entire group, is greater than 1650 m.y. A second post-Delhi intrusive period occurred at about 950 m.y. which is the age of the Erinpura Granite of the type locality.

According to Crawford, (quoted by Pichamuthu, 1971) whole-rock Rb-Sr isochron ages of post-Delhi Bairath granite and Bagan River granite are strikingly concordant (1660 m.y) and so sedimentation of Delhis must be much earlier than 1660 m.y. Sarkar, however considers that this is contradictory to other data available from Delhi and Aravalli rocks; and in his opinion the Delhi orogeny and metamorphism closed between 724 and 860 m.y. (Sarkar, 1968, p.15) and a metamorphic event took place at about 600 m.y. as indicated by K-Ar ages of biotite schist (621 m.y.) from Khetri; phyllite (643 m.y) from Babai, biotite (614 m.y) from Delhi schist, and Pb-isotopic age of the Samarskite (580 m.y) from the pegmatites of Kishangarh. Malani rhyolites occupy a large area in and around Jodhpur. The associated Jalor and Siwana granite intrusives into Delhis are considered to be later than the Erinpura granite. K-Ar ages indicate that the rhyolite flow and granite emplacement took place about 600.780 m.y. ago (Pichamuthu, 1971).

4A.4 Temperature Measurements

Temperature measurements at depth intervals of 5 or 10 m (inclined) were made in 11 boreholes: three at Madan-Kudan, two each at Kolihan and Bhagoni, one each at Kalapahar, Zawar, Pur-Doriba and Rajpura-Doriba. Vertical depths of boreholes varied from 200 m to 700 m. Particulars of these boreholes alongwith leastsquare temperature gradients calculated for depth intervals, considered to be free from disturbances as revealed by linear temperature - depth curves, are given in table IV A-1. The temperature gradients are lowest at Rajpura-Dariba and Pur-Dariba (Pre-Aravallis), somewhat higher at Zawar (Aravallis) and highest in the Delhis. Temperature depth curves in respect of ten boreholes are shown in Figures IV A-2, 3 and 4.

4A.5 Coefficient Of Thermal Conductivity

The conductivity values were determined from measurements on 249 water saturated circular discs of either 41.2 mm or 25.4 mm diameter using the Modified Benfield or the Modified Birch Apparatus (Section 2.1.2), in which a quartz disc cut both parallel and perpendicular to its optic axis or a calibrated lexan disc served as a reference material. The overall conductivity values are considered to be accurate to ± 5%.

Most of the boreholes except that at Madan-Kudan are inclined boreholes drilled at an angle so as to penetrate formations perpendicular to their plane of foliation. However

TABLE IVA-1

Particulars of Boreholes Logged, Geothermal Gradients and Heat Flow Data in Aravalli Mountain Belt

Location / Coordinates	Bore hole No.	Elevation above m.s.l.	Depth to which logged (metres) Vertical	Depth to which logged (metres) (Inclined)	Undisturbed period	Depth interval (metres)	Gradient °C/km	Heat Flow
1	2	3	4	5	6	7	8	9
I) Delhi Super Group								
Madan-Kudan 28°04'N 75°48'E	S-47	383	700		7 months	120-700	21.75	1.75
	S-48	381	454		3½ -do-	150-300	20.26	1.75
						300-460	22.39	1.83
	S-49	384	487		3½ -do-	150-300	20.62	1.71
						Mean heat flow value	-	1.76
Kolihan 29°59'N 75°49'E	KKBH-52	364.3	248	(393)	2 days	172-248	18.23	1.73
	KKBH-69		441	(475)	3½ days	107-255	18.00	1.76
Kalapahar	KKP-5	414	424	(580)	2 days	297-342	20.20	1.74
Bhagoni 27°17'N 76°24'E	DBBH-9	426.7	288.6	(365)	18 days	106.8-154.4	17.3	
						162.1-216.4	20.1	
						224.1-288.6	16.2	1.5 ± 0.15
	DBBH-18	419.3	279.4	(389)	21 months	74.2-192.2	21.2	
						192.2-279.4	16.6	

(contd..)

1	2	3	4	5	6	7	8	9
II) **Aravalli Super Group**								
Zawar 24°21'N 73°44'E	BM4-4	610	295.3	(454)	24 hrs.	71.6-295.3	14.5	1.05
III) **Pre-Aravalli Super Group**								
Rajpura-Dariba 24°58'N 74°8'E	DR4-35	581	275	(350)	41 hrs.	111.4-186.1	9.43	1.00
						186.1-227.7	11.83	0.89
						Mean Heat Flow Value		0.95
Pur-Dariba 25°21'N 74°82'E	PDB-11	427	204.6	(322)	11 days	120.5-204.6	12.7	0.95

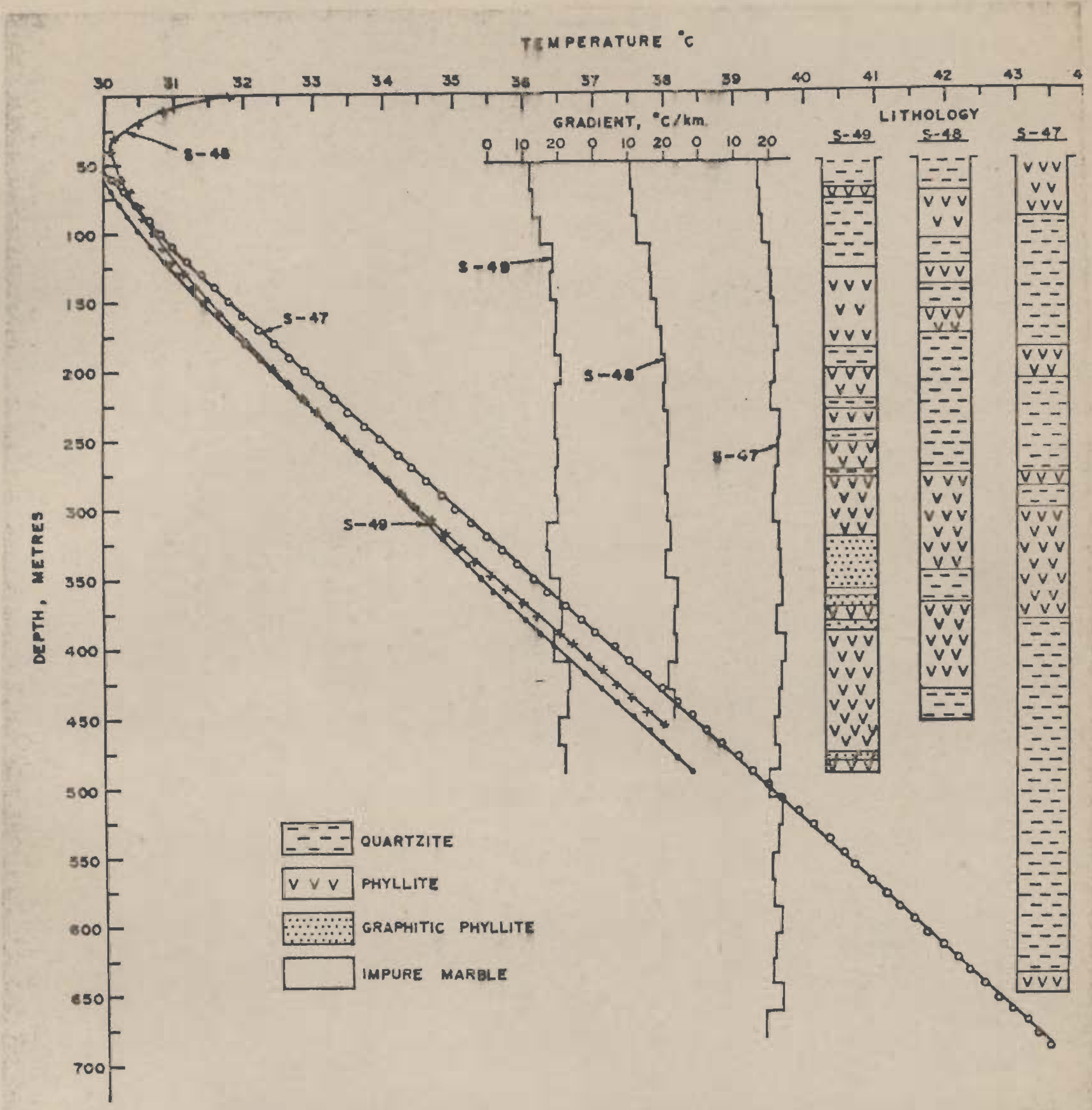


Fig. 2. Temperature versus depth curves and gradients for boreholes S-47, S-48, and S-49 in Khetri copper belt, Rajasthan.

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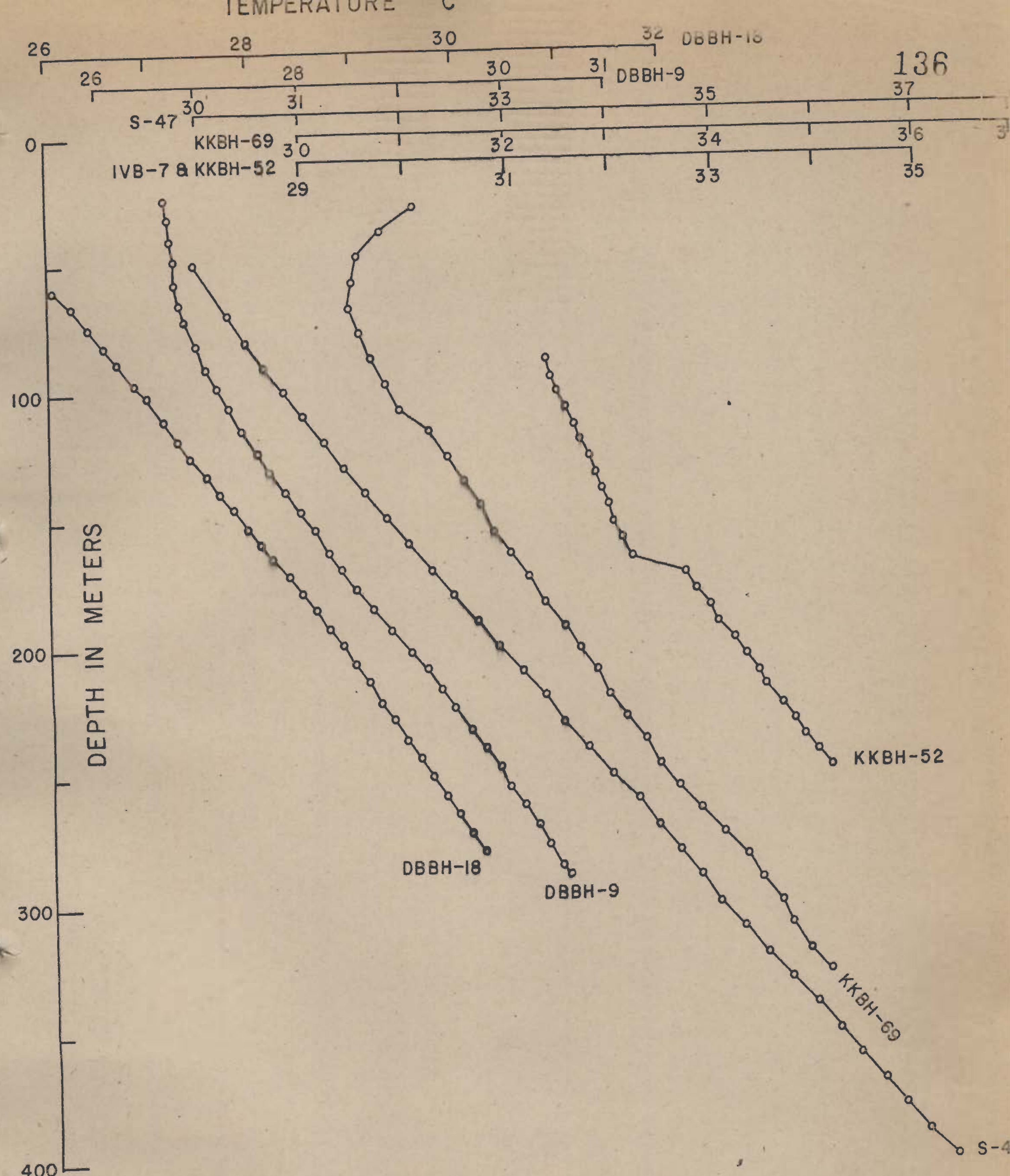

FIG. 3
TEMPERATURE PROFILES FOR BHAGONI, MADAN-KUDAN, KALAPHAR AND KOLIHAN BORE HOLES, ARAVALLI.

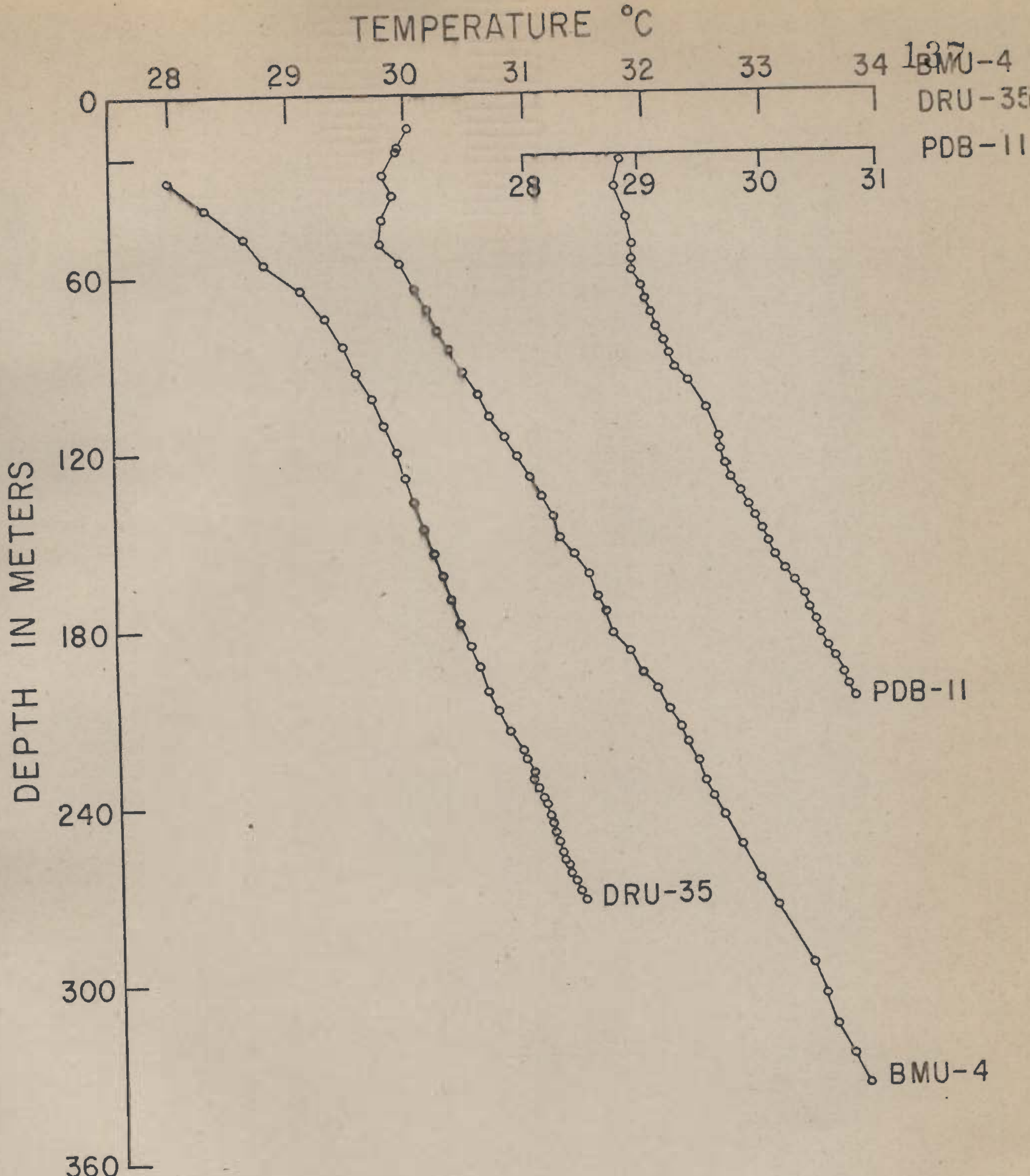

Fig. 4 Temperature profiles for Dariba-Rajpura, Pur-Dariba and Zawar bore-holes, Aravalli.

various rocks, schists in particular, show considerable anisotropy in their thermal properties. Two samples of appropriate diameters were therefore cored from each borehole and discs of suitable thicknesses prepared from these.

Let, $K_{\perp c}$ = Thermal conductivity of the sample cored perpendicular to core axis.

$K_{\parallel c}$ = Thermal conductivity of the sample cored parallel to the core axis.

$K_{\parallel f}$ = Thermal conductivity when heat flow is parallel to the plane of foliation.

$K_{\perp f}$ = Thermal conductivity when heat flow is perpendicular to the plane of foliation.

For cores taken from inclined boreholes,

$$K_{\parallel f} = K_{\perp c}$$

and

$$K_{\perp f} = K_{\parallel c}$$

Naturally the measured value of $K_{\perp c}$ would be greater than $K_{\parallel c}$, most of the time, as heat flow in such a case is mostly along the plane of foliation. Sometimes uncompatble results are obtained which may be ascribed to various factors. Thermal conductivity in the vertical direction was calculated using the relation:

$$K_v = K_{\perp c} \mathrm{Sin}^2\theta + K_{\parallel c} \mathrm{Cos}^2\theta$$

or

$$K_v = K_{\parallel f} \mathrm{Sin}^2\theta + K_{\perp f} \mathrm{Cos}^2\theta \qquad \text{IV A-1}$$

where 'θ' is the angle of dip of the plane of foliation from the horizontal.

In case there is no anisotropy,

$$K_v = K_{\perp c} = K_{\parallel c}$$

Boreholes, often deviate during drilling, causing foliation planes to make varying angles with the axis of the core sample. For such a case, using figures A and B as given below and equation (1), the following relations can be easily obtained.

$$K_{\parallel c} = K_{\parallel f} \mathrm{Sin}^2(90 - \alpha) + K_{\perp f} \mathrm{Cos}^2(90 - \alpha)$$

$$= K_{\parallel f} \mathrm{Cos}^2 \alpha + K_{\perp f} \mathrm{Sin}^2 \alpha \quad .. \quad \text{IV A-2}$$

Similarly

$$K_{\perp c} = K_{\parallel f} \mathrm{Sin}^2\alpha + K_{\perp f} \mathrm{Cos}^2 \alpha \quad .. \quad \text{IV A-3}$$

From (2) and (3), we obtain,

$$K_{\perp\!\perp f} = \frac{(K_{\perp\!\perp e} + K_{\perp e}) \cos^2\alpha - K_{\perp e}}{2 \cos^2\alpha - 1} \quad \text{.. IV A-4}$$

$$K_{\perp f} = \frac{(K_{\perp\!\perp e} + K_{\perp e}) \cos^2\alpha - K_{\perp\!\perp e}}{2 \cos^2\alpha - 1} \quad \text{.. IV A-5}$$

Thermal conductivity (K_v) of the samples in the vertical direction is then computed using the general angle of foliation of the area and the relation given in equation IV A-1.

Keeping in view the possible wide variation in the geochemical and thermo-physical properties among the various meta-sediments and between the meta-sediments and associated volcanics, an attempt has been made to make a systematic study of the thermal conductivity values of the various rock types encountered the boreholes.

Thermal conductivities of representative core samples were determined, fifty one from Madan-Kudan, thirteen from Kolihan, seventy four from Dariba-Bhagoni and twelve samples from Rajpura-Dariba and their values listed in Tables IV A-2, 3, 4, 5, 6 and 7. In order to calculate thermal conductivity in the vertical direction from the measured $K_{\perp e}$ and $K_{\parallel e}$ values, relation IV A-1 has been used for Kolihan, Rajpura-Dariba and IV A-4, 5 and 1 for Bhagoni. The samples were mostly fine-grained and were found to have negligible porosity as determined by soaking a few of these in water for 24 to 48 hours. Practically no difference was found in the conductivity values of soaked and dry samples.

TABLE IVA-2

Thermal Conductivity Values for Cores from Borehole S-47

Sample	Depth metres	Density g/cm^3	Thermal Conductivity mcal/cm sec°C	Rock Type
K-40	129	2.845	11.60	Quartzite
K-42	162	2.806	6.76	Phyllite
K-43	171	2.743	7.70	Biotitic quartzite
K-45	203	2.745	7.87	Phyllite
K-46	216	2.722	7.03	Phyllite
K-47	228	2.727	7.48	Phyllite
K-48	243	2.745	8.05	Biotitic quartzite
K-50	272	2.757	9.42	Biotitic quartzite
K-52	294	2.834	8.11	Phyllite
K-53	309	2.736	8.18	Quartzite
K-55	340	2.790	8.11	Phyllite
K-56	358	2.735	8.57	Quartzite
K-57	386	2.65	7.47	Phyllite
K-58	390	2.784	6.65	Phyllite
K-60	423	2.762	9.80	Phyllite
K-61	443	2.830	8.30	Phyllite
K-62	458	2.794	8.53	Biotitic quartzite
K-64	488	2.795	7.54	Phyllite
K-66	524	2.699	9.17	Phyllite
K-67	539	2.747	6.71	Biotitic quartzite
K-68	556	2.783	7.16	Phyllite
K-69	569	2.743	7.23	Biotitic quartzite
K-70	588	2.714	7.87	Biotitic quartzite
K-71	599	2.722	7.40	Biotitic quartzite
K-72	613	2.698	9.59	Biotitic quartzite
K-74	645	2.720	7.32	Biotitic quartzite

Conductivity range ≠ 6.65 to 11.60; mean value = 8.06 ± 1.11 s.d

TABLE IVA-3

Thermal Conductivity Values for Cores from Borehole S-48 & S-49

Sample	Depth metres	Density g/cm^3	Thermal conductivity mcal/cm sec°C	Rock Type
		Borehole S-48		
K-97	153	2.652	11.5	Quartzite
K-98	158	2.733	6.9	Biotite quartzite
K-99	186	2.63	8.1	Quartzite
K- 100	205	2.716	7.3	Biotite quartzite
K-101	213	2.689	8.4	Quartzite
K-102	229	2.659	8.3	Quartzite
K-103	245	2.735	9.2	Biotite quartzite
K-105	290	2.669	9.3	Quartzite
		Conductivity range = 6.9 - 11.5		
		Mean value = 8.63 ± 1.32 s.d		
K-107	307	2.658	7.4	Biotite quartzite
K-109	324	2.623	8.2	Quartzite
K-110	341	2.660	9.9	Biotitic quartzite
K-111	355	2.71	8.6	Biotitic quartzite
K-112	368	2.823	7.9	Biotitic quartzite
K-113	383	2.773	7.4	Phyllite
K-114	401	2.673	10.4	Biotitic quartzite
K-115	419	2.784	8.7	Phyllite
K-116	434	2.84	7.9	Phyllite
		Conductivity range = 5.4 to 10.5		
		Mean value = 8.18 ± 1.39 s.d		

(Contd..)

T A B L E IVA-3 (contd..)

Sample	Depth metres	Density g/cm^3	Thermal Conductivity mcal/cm sec °C	Rock Type
		Borehole S-49		
K-7	157	2.749	7.60	Phyllite
K-8	166	2.728	7.40	Phyllite
K-10	195	2.752	7.90	Biotitic quartzite
K-11	216	2.751	7.90	Biotitic quartzite
K-12	236	2.71	11.50	Quartzite
K-14	272	2.819	7.80	Quartzite
K-15	292	2.770	7.45	Phyllite
K-16	299	2.847	8.90	Phyllite

Conductivity range = 7.4 to 11.5

Mean value = 8.31 ± 1.37 s.d

TABLE IVA-4

Thermal Conductivity Values for Cores
from Kolihan Boreholes KKBH-52 and KKBH-69.

Vertical depth metres	KKBH-52 Thermal Conductivity*			Vertical depth metres	KKBH-69 Thermal conductivity*		
	K_c	K_c	K_v		$K_{\perp c}$	$K_{\parallel c}$	K_v
177	11.72	11.69	11.70	136	12.22	10.07	11.97
183.5	8.35	6.32	7.16	145	11.2	8.54	10.69
190	6.15	5.5	8.20	174	8.45	6.8	8.25
196.5	10.51	11.65	11.18	201	8.9	6.5	8.62
203	11.6	9.6	10.43	210	9.0	8.76	9.0
209.5	10.4	8.5	9:28	219	10.2	8.73	10.03
215.7	11.75	10.34	10.92				
Mean value	10.07	9.09	9.84	Mean value	10.71	8.61	9.79

*mcal/cm sec°C.

T A B L E IVA-5

Thermal Conductivity Values of Core Samples from Borehole DBBH-9.

Rock Type	Inclined Depth metres	Thermal Conductivity mcal/cm. Sec°C $K_{\parallel c}$	$K_{\perp c}$	K_{v}
Arkosic Quartzite	177.3	11.66	12.9	12.1
	187.5	10.96	9.12	10.3
	197.6	9.31	10.20	9.6
	207.3	9.85	11.34	10.4
Mean value		10.45	10.89	10.6
Biotitic quartzite	287.1	9.44	10.12	9.7
	294.3	10.49	10.83	10.6
	301.9	7.27	7.66	7.4
	308.9	9.63	9.58	9.6
	316.0	9.51	9.99	9.7
Mean value		9.27	9.49	9.4
Amphibole Quartzite	323.0	7.22	6.15	6.8
	330.6	9.00	8.52	8.8
	336.8	8.67	7.75	8.3
Mean value		8.3	7.44	8.0
Metadelerite or Amphibolite	217.4	6.00	6.24	6.1
	227.5	6.00	6.02	6.0
	237.6	6.24	6.00	6.1
	257.5	6.97	6.59	6.8
	267.5	6.86	6.72	6.4
	352.3	5.95	5.60	5.8
	358.9	5.96	5.79	5.9
	365.3	6.79	6.45	6.7
Mean value		6.35	6.18	6.2
Phyllite	129.8	9.40	10.07	9.6
	133.8	8.17	9.72	8.7
	167.4	7.34	10.44	8.5
	168.0	7.89	10.15	8.7
Mean value		8.20	10.10	8.9

T A B L E IVA-6

Thermal Conductivity Values of Core Samples from Borehole DBBH-9

Rock Type	Inclined Depth metres	Thermal Conductivity mcal/cm Sec°C $K_{\parallel c}$	$K_{\perp c}$	K_{v}
Metadolerite and Amphibolite	230.0	7.4	7.8	7.5
	235.1	6.8	6.8	6.8
	240.0	5.4	5.5	5.4
	250.0	6.8	6.8	6.8
	255.0	6.7	7.0	6.8
	260.0	6.45	6.7	6.5
	265.9	7.4	7.0	7.2
	270.0	8.91	6.9	8.2
	275.4	6.8	7.7	7.1
Mean value		6.96	6.91	6.92
Phyllite	100.8	6.7	9.3	7.7
	120.4	7.3	8.4	7.7
	175.2	8.8	10.7	9.5
Mean value		7.6	9.47	8.3
Marble	140.2	5.7	5.7	5.7
	206.0	-	5.5	5.5
Mean value			5.6	5.6
Amphibole Quartzite	245.6	6.2	6.1	6.16
	253.3	8.1	8.8	8.4
Mean value		7.15	7.45	7.28
Quartzite (Arkoric, Felspaltic or Gritty)	185.7	11.1	10.6	10.9
	190.0	9.5	10.4	9.8
	194.9	11.5	11.4	11.5
	210.0	9.4	8.5	9.1
	215.1	8.8	9.7	9.1
	220.0	9.6	11.1	10.1
	225.8	9.6	10.5	9.9
Mean value		9.93	10.31	10.1
Biotite Quartzite	285.5	9.3	9.8	9.5
	305.4	10.2	9.3	9.9
	365.8	10.1	10.4	10.2
Mean value		9.87	9.83	9.87

TABLE IVA-7

Thermal Conductivity Data for Rajpur-Doriba borehole DR4 - 35

Rock Type	Vertical depth metres	Thermal Conductivity* $K_{\perp c}$	$K_{\parallel c}$	K_v	$K_{\perp}/K_{\parallel c}$
Kyanite-staurolite bearing Garnetiferous graphitic schist	101	12.32	7.46	9.14	1.65
-do-	107	12.5	9.34	10.43	1.34
-do-	126	13.2	11.28	11.94	1.17
-do-	133	13.6	8.69	10.39	1.21
-do-	143	11.0	8.42	9.31	1.31
-do-	144	13.68	11.73	12.40	1.17
Mean value		12.72	9.49	10.60	
Graphitic mica schist	180	9.5	6.6	7.6	1.44
-do-	219	10.55	4.52	6.67	2.28
-do-	220	10.51	6.08	7.61	1.73
-do-	224	11.15	5.52	7.46	2.02
-do-	229	10.16	5.62	7.19	1.81
-do-	234	12.46	6.35	8.46	1.96
Mean value		1072	5.8	7.5	

*mcal/cm sec°C

4A.6 Evaluation of Heat Flow

4A.6.1 Heat flow in the Delhi super group formations

Thermal measurements have been carried out in three areas of the Khetri Copper Belt and one of the Alwar Copper Belt of this Super Group. The Khetri copper belt which is about 40 km long is situated at the northern end of the Aravalli system. The rock formations in the area consists mostly of quartzites and phyllites belonging to the Alwar and the Ajabgarh series of the Delhi system. Regionally the strike of the formations varies from NNE-SSW to NE-SW with high dips.

Widely distributed copper occurrences have been found in the Alwar and Raialo series of the Delhi Super Group near Alwar town, mineralisation being confined primarily to the calcareous rocks of the Upper Raialo Series in Bhagoni where thermal measurements have been made in two boreholes DBBH-9 and 16 (Table IV A-1).

4A.6.1.1 Results from Madan-Kudan

Temperature versus depth curves for three bore holes along with gradients at different depths are shown in Figure IV A-2. Since the temperature - depth curve for borehole S-47 is quite linear below a depth of 120 m, only one gradient was calculated for the interval 120-700 m. The least-square fit of temperature with depth for this borehole is T(°C) = 28.5 + (21.75 ± 0.03) X, where 'X' is depth in kilometres.

In borehole S-48, two slightly different gradients were found in the depth intervals, 150-300 m, and 300-460 m. These values were 20-26 ± 0.08°C/km and 22.39 ± 0.09°C/km, respectively. The overall gradient in 150 to 460 metre depth interval being 21.28 ± 0.1°C/km.

In borehole S-49 three distinctly different gradients were found in the depth intervals 130-300 m, 310-370 m, and 380-500 metres. The three gradients in S-49 were found to be 20.62 ± 0.09°C/km, 18.33 ± 0.4°C/km, and 21.74 ± 0.2°C/km respectively. The change in gradient below 310 m may be due to the presence of graphitic phyllites lying between 320 and 370 m.

Heat flow values evaluated from these boreholes for four different depth intervals are listed in Table IV A-1. For borehole S-49, heat flow was evaluated for only one interval because adequate number of samples were not available at other depths.

The mean value of heat flow obtained from the three boreholes is 1.76 ± 0.1 HFU. Differences in heat flow values may be due to inadeqquate samples for thermal conductivity measurements.

4A.6.1.2 Results from Kolihan and Kalapahar

These two sections of the Khetri copper belt (KCP) are located within ten kilometres from Madan-Kudan. Observed temperature gradients for these two sections of KCP, are slightly lower than those observed in Madan-Kudan. However, the heat flow values are more or less of the same order (Table IV A-1). The heat flow value from borehole KKBH-69 relates to only one depth interval (107 - 255 m), because core samples could not be obtained from deeper sections of the borehole. No core samples could be obtained from borehole KKP-5 either, but fortunately its detailed litholog was available. This and the conductivity values of rock types of interest, given in tables IV A-2, 3 and 4, were used for heat flow evaluation in respect of borehole KKP-5.

4A.6.1.3 Results from Bhagoni

Apart from various types of quartzites and phyllites, thick sections of metadolerite and amphibolite also occur in Bhagoni area. Coefficients of thermal conductivity of these rocks and quartzites are quite different. Two methods viz., i) KG and ii) MBP (sections 2.3.1 and 2.3.3 respectively) were used for the evaluation of heat flow. The conductivity of the homogeneous rock underlying the whole structure in which heat flow would be normal was taken as 7.5 μcal/cm.sec.°C. The plots of temperature against $Di\left(\frac{\cos^2\theta}{Ki} + \frac{\sin^2\theta}{Ko}\right)$ in respect of both the boreholes, gave two linear segments each. The heat flow values are given in Table IV A-8. It appears that the effect of refraction due to dipping beds of different conductivity is rather appreciable and the method of correction has not proved very satisfactory in the present case.

The heat flow at Bhagoni (1.5 $\pm$ 0.15 HFU) although lower than that for the north-eastern part of the Khetri copper belt, is still higher than the global average for shield areas (Chapter IV C).

4A.6.2 Heat flow in the Aravalli system

Thermal measurements were carried out in borehole No. BM4-4 drilled in the rocks of Aravalli Super Group in the Zewar lead-Zinc belt, which extends along a distance of about 20 km.

The boreholes crossed through rock formations consisting mainly of graywake with some phyllite. Although core samples of rocks encountered in the borehole could not be obtained, a heat flow value of 1.05 HFU was obtained using the relationship between surface heat flow and average temperature gradient for the Pre-Cambrian crystalline areas of India (Figure IV A-5).

4A.6.3 Heat flow in Pre-Aravalli super group

Heat flow values in this Super Group have been determined from thermal measurements at two locations in the Bhilwara Lead-Zinc-Copper belt, which extends along a distance of about 80 km from Rajpura Dariba to Banera (Figure IV A-1).

TABLE IV A-8

Surface Heat Flow in Bhagoni

Bore hole No.	Depth interval metres	Thermal conductivity m cal/ cm. sec°C	No. of sam-	Gradient. °C/km	Heat Flow µcal/cm²sec. Q**	Q***
DBBH-9	106.8-154.5* (130 - 190)	9.65	6	17.3	1.67	
DBBH-9	162.1-216.4 (200 - 270)	7.00	6	20.1	1.40	1.47
DBBH-9	224.1-288.6 (280 - 365)	8.27	10	16.2	1.34	1.31
		Mean heat flow value for 3 depth intervals			1.47	
DBBH-18	74.2-192.2 (100 - 270)	8.52	23	21.2	1.67	1.45
DBBH-18	192.2-279.4	-	-	16.6	-	1.32

*Inclined depth; **by KG Method; ***by MBP Method;

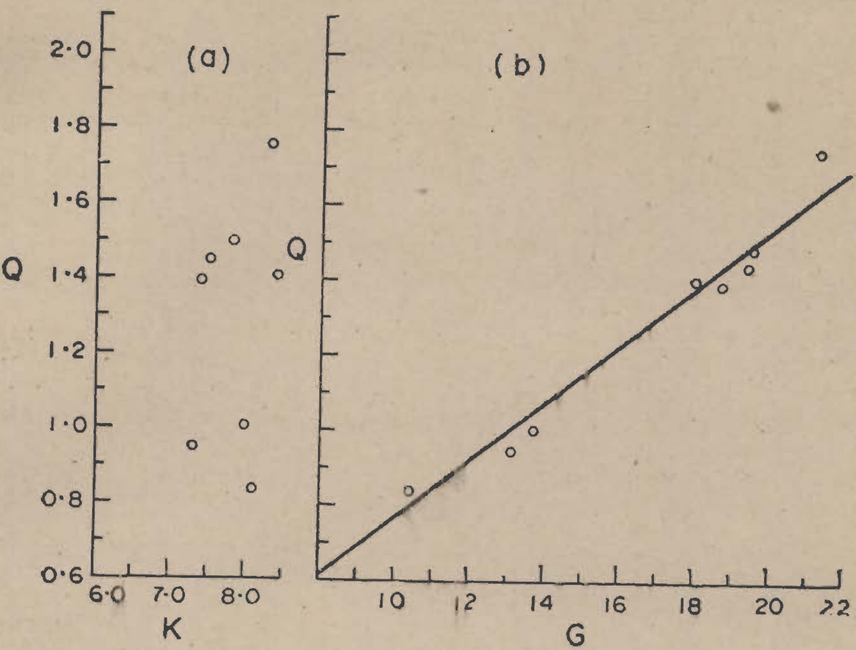

FIG. 5—(*a*) Heat flow Q (in μ cal/cm^2 sec), versus thermal conductivity K (in m cal/cm sec °C).

(*b*) Heat flow Q versus geothermal gradients, G (in °C/km) for Pre-Cambrian crystalline areas in India.

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The Rajpura-Dariba area constitutes the western limb of a megasyncline and is underlain by high grade metamorphic rocks. Much of the area, however, is covered by alluvial.

The sulphide mineralisation is mainly confined to graphitic mica schist at Rajpura-Dariba and banded magnetite-quartzites which is interlined with quartz-sericite schist at Pur-Dariba.

Least square gradients and heat flow values at Rajpura-Dariba were evaluated for two vertical depth intervals 111.4 to 186.1 m and 186.1 to 227.7 m. In the first interval, the gradient was found to be 9.43°C/km which considering a mean conductivity value of 10.6 μ cal./cm.sec.°C (Table IV A-7) yields a heat flow value of 1.00 HFU. The gradient at deeper interval is higher than that observed in the shallow depth interval, but as the rocks at this level has a lower conductivity value of the heat flow value turns out to be 0.89 HFU. Observed variation in the heat flow values appears to be caused by unsystematic conductivity variations coupled with some effect of refraction in dipping strate of different conductivities. The mean heat flow value obtained from these two depth intervals is 0.95 HFU.

Least square gradient and heat flow values were similarly calculated in respect of borehole PDB-11 for one vertical depth interval of 120.5 to 204.6 m. The gradient in this depth interval is 12.7°C/km, which considering the mean conductivity value of 7.51 mcal/cm.sec°C obtained from 6 samples yields a heat flow value of 0.95 HFU.

Discussion of Results

The heat flow data as obtained for the Aravalli mountain system clearly demonstrates that the NE part of the Khetri copper belt is characterised with significantly higher heat flow than other parts of the belt. Another point that emerges clearly is that heat flow over areas covered with rocks of the Delhi Super Group is much higher than over rocks of the Aravalli and Pre-Aravalli Super Groups. In the latter two super groups, the heat flow values agree with the characteristic trend for Precambrian Shields. These two super groups x are more or less equivalent to the Dharwars, heat flow values over which are also similarly low (section 4A.7 and Chapter V B). A detailed discussion of the high heat flow values is given in chapter IV C.

4A.7 Heat Flow At Kalyadi, Dharwars

4A.7.1 General geological setting

Kalyadi (13°04'N; 76°9'E, Figure IV A-6) is located on a sharply outlined northwest plunging synclinorium of schistose rocks known as Dharwars which runs in the NNW-SSE direction amidst the vast expanse of Peninsular Gneisses near a patch of granite called the Arsikeri granite. It forms an isolated hill range of about 8 miles in length and 5 miles in

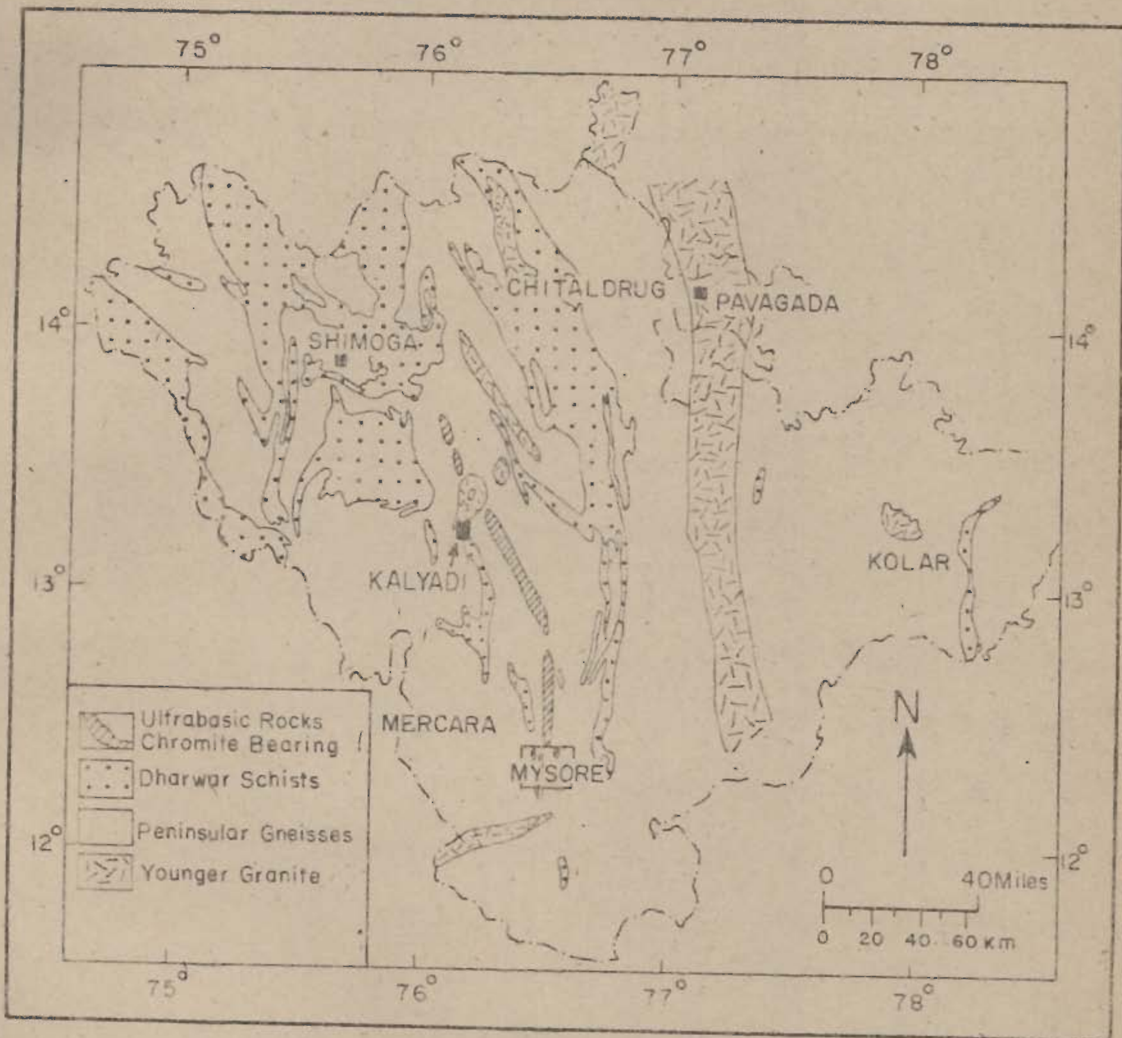


FIG. IV-6 Geological map of the area around Kalyadi
(After Geological Survey of India)

IV -35

Fig 5

width and lies between Latitude 13°19'N - 13°27'N; and Longitude 76°15'E - 76°19'E. Studies show that it did not crystallize out of a liqquid magma but developed insitu (Das, 1970). The Dharwar formations, representing one billion years of Pre-Cambrian era (3.2 - 2.2 b.y) are mainly made up of metasediments and metavolcanics, which have been folded cross-folded, metamorphosed and intruded by various granites and dykes. These rocks are exposed in NNW-SSE, N-S and NNE-SSW trending narrow belts. Some of the oldest dates in India have been reported in respect of rocks belonging to the Dharwar schists belts, viz., a K-Ar age of 3295±200 m.y. for hornblende schist from Hutti Gold (Sarkar, 1968) and lead isochron af ages of 2900±200 m.y. for galena from Chitaldrug (Vinogradov and Tnagrinov, 1964). The prominent rock type in the Kalyadi area is quartzite biotite-chlorite-schists, and biotite gneiss and ultrabasic rocks viz., talc-chlorite and Actinolite-chlorite-schist, pink porhyritic granites and Pegmatites. Mineralization in the form of pyrite and chlcopyrite is confined to both schists and sheared quartzites which are interbedded. The country rock is a grey-banded granite gneiss which is highly crushed and micaceous (Radhakrishna, 1974).

4A.7.2 Temperature measurements

Temperatures were measured in an inclined borehole KAL-H2 (45°) at 5 or 10 m intervals within a depth interval of 270-350 m. The temperature depth curve and the gradients in the borehole are shown in Figure IV A-7.

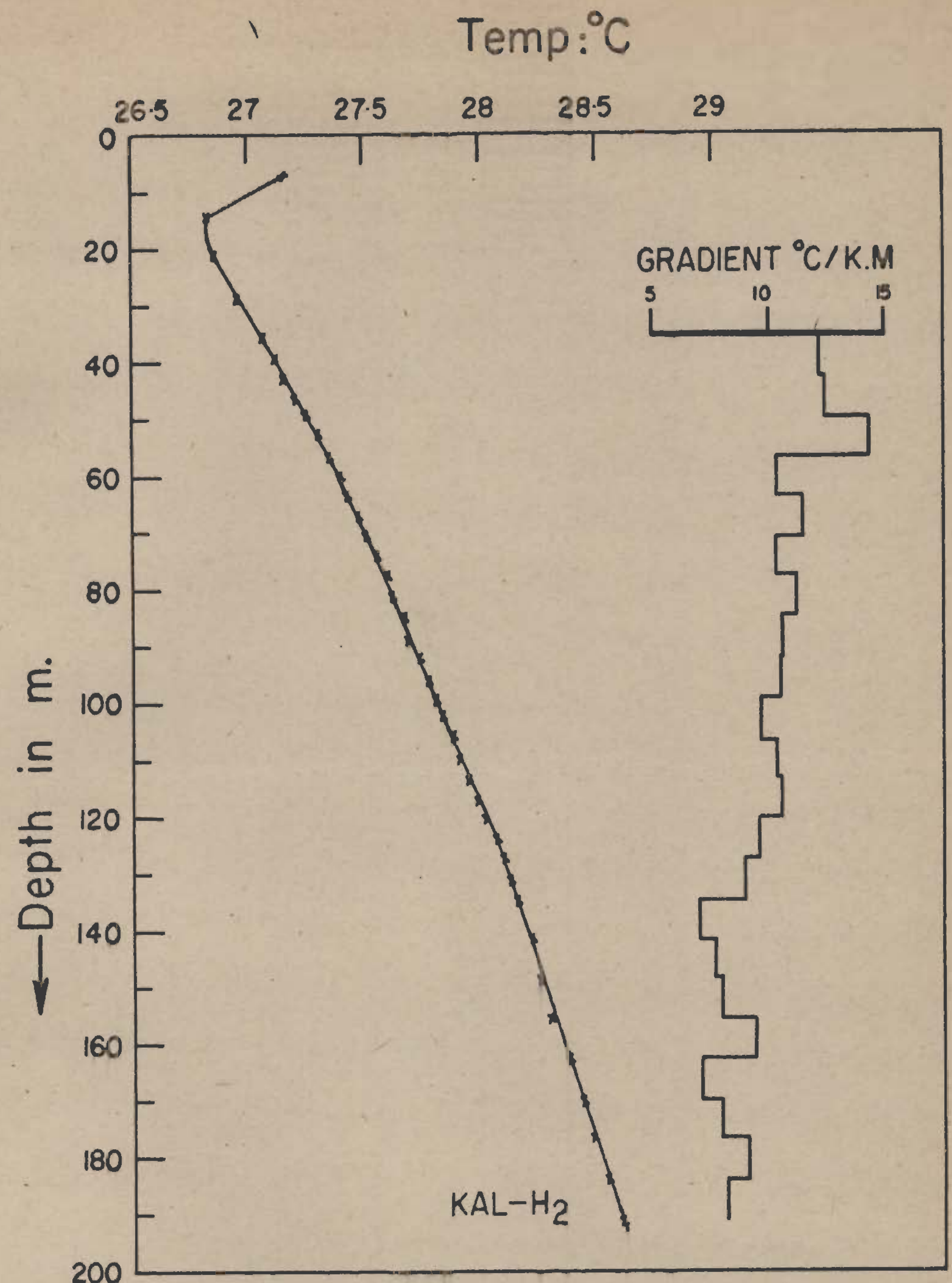


FIG. IV-7 Temprature versus depth curve and gradients for bore hole KAL-H_2 in Kalyadi(Dharwars)

4A.7.3 Thermal conductivity measurements

Coefficients of thermal conductivity of 15 core samples collected from the borehole were measured, using the Modified Birchs apparatus. Sample discs were cored both parallel and perpendicular to the core axis. From these values, conductivities were then calculated for the plane of Schistocity ($K_{\parallel}$), perpendicular to the plane of schistocity ($K_{\perp}$), and finally in the vertical direction (K_v), following the method stated earlier (Page IV A-18).

4A.7.4 Evaluation of heat flow

Least square gradients and heat flow values were evaluated for two vertical depth intervals: 81.3 to 130.8 m and 134.3 to 190.9 m. In the first interval the gradient was found to be 10.21°C/km. Combined with the weighted mean conductivity of formations (7.83 m cal/cm^{2}sec.°C), the heat flow turns out to be 0.8 HFU. The gradient in the deeper interval where mineralised quartzite has been encountered was found to be 8.0°C/km which is lower than that observed in the upper depth interval. Lowering of the gradient is mainly caused by the higher conductivity values of these quartzites as compared with those for schists which alongwith pegmatites occur in the upper depth interval. In the second interval the weighted thermal conductivity was 8.0 mcal/cm^{2}sec°C and the corresponding heat flow value equal to 0.64 HFU. The observed variation in heat flow is mainly due to unsystematic conductivity variations. The formations in the depth interval 176.8 to 190.9 m show occassional specks of pyrite. This depth section when considered separately, yielded a heat flow value of 0.67 HFU, the mean heat flow value from the two depth intervals being 0.74 HFU.

4A.8 Discussion of Resulsts

h.f.u

The heat flow value of 0.74 for Kalyadi is typical of a shield area. In the southern part of the Indian shield one other value of heat flow viz., 0.95 HFU at Kolar in the Dharwar schist belt has been reported (Rao, 1970). The Kolar as well as the Kalyadi schist belts, are both eq uivalent to the green stone bel ts of other shields which are also characterized by low heat flow. Rocks of both the areas show high grade of metamorphism. The foliated biotite gneiss which was met betWeen the 184-191.2 m vertical depth interval in the Kalyadi borehole contained pink garnet.

CHAPTER IV B

CRUSTAL THERMAL STRUCTURE OF DHARWAR CRATON

4B.1 Introduction

The Dharwar Craton iscomposed of various systems of Pre-Cambrian rocks, evolved through complex geological processes. It has remained a stable land mass for more than 600 m.y. except for some gentle warping. Some values of surface heat flow and heat generation in the near surface rocks of the Dharwar Craton have been reported earlier (Gupta and Rao, 1970; Rao and Rao, 1975; Atal et al. 1978). These together with some more new data recently obtained have been used to estimate the crustal temperatures of a few units of the Dharwar Craton.

4B.2 Pre-Cambrian Systems of the Dharwar Craton

According to Narayanswamy (1970), three major rock systems of this Craton are enclosed within the Peninsular Gneiss basement complex. These are:

1) The Dharwar System - schist belts
2) The Chanrnokite - khondalite system
3) The Cuddapah - Kurnool system.

Without going into the controvercial relationship of the Dharwar sonist belts with the gneissic complex, one may summarise the salient features of the Dharwar system as follows:

4B.2.1 Dharwar system

The Dharwars consist of the famous schistose predominantly metasediments and metavolcanics rocks of India exposed in NNW-SSE, N-S and NNE-SSW trending narrow belts. The prominent schist belts are those of Shimoga-Goa, Chitradurga-Gadag. Bababudan, Kolar, Sandur, Holenarasipur, Nuggihalli etc., (Figure IV B-1). It is generally recognised that while some schist belts viz., the Kolar, Sandur, Hutti etc., are essentially volcanic, similar to other well known greenstone belts of the world, others such as the Shimoga-Goa, the Chitradurga-Gadag and the Bababudan belts are predominantly sedimentary with intercalated volcanics (Swaminath et al., 1976; Naqvi, 1977). Keeping these features in view, Swaminath et al. (1976), suggested a possible division of the Karnataka Craton, which is a part of the main Dharwar Craton into an Eastern Block dominated by volcanic greenstone belts, and a Western Block characterised by sedimentary greenstone belts, roughly separated by the Closepet Granite (Figure IV B-1).

Almost all the schist belts which are widely distributed, are associated with positive gravity anomalies varying in amplitude from 10 to 30 mgal. In most cases the gravity contours are found to follow the configuration of the schist belt with localised negative anomalies over the intruding granites (Subrahmanyam, 1978). The thickness of the schist belts as yielded by gravity data and by deep seismic sounding

FIG. IV-B1 Greenstone belts and high grade schists of Dharwar Craton (After Swaminath et al.1976) studies respectively range from about 4 to 10 km (Subrahmanyam, Personal communication) and 4.5 to 7.5 km, (Kaila et al. 1979).

4B.3 Heat Flow And Heat Generation Relationship

Heier (1965) and Lambert and Heiër (1968) have shown that processes of regional metamorphism, partial melting and magmatism which ceased long ago in the Precambrian cratons, cause radioactive heat producing elements to migrate upwards thereby increasing their concentration at higher levels in the crust.

Birch et al. (1968) and Lachenbruch (1968) established that in many Pre-Caledonian regions of USA, the surface heat flow Q and heat generation of the surface rocks 'A_o' have the following linear relation.

$$Q = Q_o + A_s\ b \qquad \text{... IV B-1}$$

where Q_o and 'b' are constants with dimensions of heat flow and length (depth) respectively, and are characteristic of large provinces of uniform tectonic structure. The value of 'b' obviously represents the slope of the line between Q and A_o.

The above mentioned relation led to the division of the surface heat flow into two components; one connected with a top radioactive layer and the other 'Q_o' (the reduced heat flow) upflowing from the lower crust and mantle. The relation also provided of clear idea of the vertical distribution of heat production and more precise estimate of crustal temperatures. There are innumerable solutions for distribution of heat generation versus depth which can satisfy the above mentioned emperical relationship. However, two extreme distributions that have been much discussed (Birch et al. 1968; Roy et al. 1968; Hyndman et al. 1968; Lachenbruch, 1968) are: i) assumption of a constant heat generating layer upto a depth 'b' (the step model) and ii) heat production decreasing exponentially with a logarithmic decrement 'b' according to the relation

$$A(A) = A_o e^{-Z/b} \qquad \text{... IV B-2}$$

where Z is the depth.

The step model represents the simplest interpretation of the above relation and can easily account for lateral variations in the surface heat flow over a province, due to horizontal variations in heat generation in a top radioactive layer of thickness 'b'. In respect of the exponential model the layer subject to horizontal variation in heat production can be of any thickness.

4B.4 Geothermal Data And Discussions

The Dharwar system of rocks have been fairly well studies owing to their associated accessories deposite.

Various exploratory boreholes have been drilled in the schist belts and reliable heat flow values have been obtained at seven locations. However, the Peninsular Gneisses although covering a vast terrain of the Dharwar Craton have induced comparatively lesser attention being largely barren, and only a few boreholes exist there. Thermal logging of one of these was carried out upto a depth of about 200 m. The mean heat flow values and available data on mean heat generation in various rocks of the Dharwar Craton are summarised in Tables IV B-1 and IV B-2 respectively.

Heat flow within the Dharwar Craton generally varies from 0.60 to 1.31 HFU with a solitary high value of 1.8 HFU, observed at the NE margin of the Cuddapah basin. We also notice that the Dharwar Schist belts of the western block (Table IV B-1 and Figure IV B-1) are characterised with distinctly lower heat flow than the schist belts of the eastern block. The lowest heat flow value (0.6 HFU) free from any concievable disturbance was obtained from a fairly deep bore hole in the Dharwar Schist belt of Western block which based on gravity and deep seismic data has a thickness of about 7 km. Observation of this low heat flow in the Dharwar Craton imposes a constraint on the magnitude of 'Q_0' (equation 1). Even if we consider a lower value of 0.6 HFU for A_0 (Table IV B-2), and a thickness of 7-8 km for thetop radiactive layer, a value of about 0.55 HFU is obtained for 'Q_0'. An attempt has been made to obtain the values of the two heat flow parameters 'Q' and 'b' for the Karnataka Craton (Figure

TABLE IV B-1

HEAT FLOW DATA IN DHARWAR CRATON

S.No.	Description	Depth of boreholes (m)	No. of values	Mean value 10^{-6}cal. cm^{2}.sec
1.	Dharwar Schist Belts			
a)	Sedimentary Greenstone Belts (Western Block)	150 to 630	4	0.64
b)	Volcanic Greenstone Belts (Eastern Block)	150 to 165 deep mines (Kolar)	2 1	0.93
2.	Peninsular Gneisses			
	Karadikuttam, Madurai district.	195	1	1.31
3.	Cuddapah Basin			
a)	North-eastern margin (Bandalamattu)	800	1	1.80
b)	South-western margin (Pulivendala)	230	1	0.64

TABLE IV B-2

SUMMARY OF HEAT GENERATION DATA

S. No.	Area and System	Rock Type	U ppm	Th ppm	K%	Heat generation HGU
1.	Dharwars					
	a) Holenarasipur Schist Belt	i) anorthosite	0.2	0.5	0.06	0.24
	-do-	ii) amphibolite	0.4	1.1	0.19	0.55
	-do-	iii) granitic gneiss	2.8	7.5	1.20	3.30
	b) Kolar Schist Belt	i) amphibolite	0.4	1.3	0.20	0.62
	-do-	ii) hornblende schist	-	-	-	0.62
	-do-	iii) granitic gneiss	-	-	-	3.60
	c) Sargur Schist Belt	amphibolite	0.25	0.8	0.20	0.38
2.	Peninsular Gneisses					
	a) Kardikuttam (Madurai district)	gneisses	-	-	-	7.00
	b) S.E. of Mysore	Banded gneisses	-	-	-	7.30
3.	Cuddapah Basin					
	Bandalmattu	Phyllites and argillites	-	-	-	6.25

1, HGU = 10^{-13} cal/cm^{3}sec = 0.4187×10^{-6} Wm^{-3}

IV B-1) by using three data sets of heat flow and heat generation in near surface rocks (Tables IV B-1 and 2). Values of 0.54 and 11.3 for 'Q_o' and 'b' respectively have been obtained. On account of varying thicknesses of the schist belts and therefore of the top layers, these can be taken as approximate values only. However, it is quite evident x that a major part of the Dharwar Craton is characterised by low heat flow from the lower crust and upper mantle.

4B.5 Crustal Temperatures

Jaeger's (1965) equation was used to estimate crustal temperatures. The equations given below relate temperature to the surface heat flow and subsurface heat source in a layered crust, for steady flow of heat in one dimension, we have

$$T(Z) = T_o + \frac{QZ}{K} - \frac{A_o Z^2}{2K} \qquad \text{... IV B-3}$$

$$T(Z) = T_o + \frac{QZ_1}{K} - \frac{A_o Z_1^2}{2K}$$

$$+ (Q - A_o Z_1)(Z - Z_1)/K_1$$

$$- A_1 (Z - Z_1)^2/2K_1 \qquad \text{... IV B-4}$$

Where T_o is the mean surface temperature and Q the surface heat flow. The equations refer to layers of constant heat production and thermal conductivity. Equation 3 is for a top layer of radioactive heat generation A_o and thermal conductivity K. When these values change abruptly to A_1 and K_1 at depth Z_1 the temperatures for depths Z and Z_1 are given by equation 4.

Subsurface temperatures for a crustal model in which heat production decreases exponentially with depth, are given by the following expression (Lachenbruch, 1968, 1970).

$$T(Z) = T_o + \frac{Q_o Z}{K} + \frac{A_o b^2 (1 - e^{-Z/b})}{K} \quad \ldots \text{ IV B-5}$$

Values of average heat production, in the lower crustal rocks, knowledge of which is uncertain, is required for calculations of temperature using equation (4). Geochemical deductions do exist and a constraint is also imposed on the value of A_1 by Q_o. However, it is believed that the lower crust is composed of high grade basic rocks. Heat production value in high grade metamorphic rocks ranges from 1.1 to 1.3 HFU, Lambert and Heier (1967), and is 1.08 in average basalts, Heier and Rogers (1963). These values lead to very low mantle heat flow 'Q_m' (0.17 to 0.24 HFU) in the present case and therefore appear unacceptable. It is most probable that the lower parts of the crust, specially under the Karnataka Craton, was depleted in radioactivity, and the present heat generation is of the order of 0.5 HGU. On similar considerations Blackwell (1970) adopted almost the same heat generation value for his model of Sierra Nevada, Allis (1979), while proposing a heat production model for a stable continental crust, pointed out that in such a case the lower crust should be uniformly depleted of heat producing elements with a mean heat production of 0.5 ± 0.25 HGU.

A coast to coast, deep seismic sounding profile across the Dharwar Craton, has shown that the crustal thickness varies from about 34 to 40 km (Kaila et al. 1979). Accordingly a value of 38 km has been adopted as the average crustal thickness of the Dharwar Craton. An estimate was then made of the surface heat flow component from the mantle (Q_m) underlying the Dharwar Craton, assuming the calculated value of Q_o = 0.55 HFU and b = 11 km which turns out to be 0.41 HFU. This value is similar to those inferred by Hyndman and Everett (1968) for the mantle underlying the Kambalda x region of the West Australian shield. Whilst the magnitude of 'Q_m' is likely to differ from one shield to another (Gupta et al. 1971; 1975), a constant mantle contribution to the surface heat flow is largely favoured from individual Pre-Cambrian Craton. On this assumption we shall examine its implications to crustal temperatures of the Dharwar Craton.

The diverse types of surface geology for which geothermal data (Tables 1 and 2) is available are: i) the greenstones; ii) the Peninsular Gneisses and iii) the Proterozoic metasediments. Crustal temperature have been calculated for a number of models incorporating different values of heat production and thicknesses. The parameters assumed for such models are listed in Table IV B-3 and the resulting temperatures in Table IV B-4. Models G_1 and P_1 pertain to a two layer crust with the top radioactive layer having a thickness of 11 km while models G_2 and P_2 represent exponentially decreasing values of heat generation. Using the available seismic information, a three layer model 'P_3' (Table IV B-3) has also been considered. The velocity-depth function as determined from the Deep Seismic Sounding data in the Dharwar Craton has shown that the interval velocity starting from a value of 5.68 km/sec at a depth of 2 km, increases continuously and reaches a value of 6.5 km/sec at 15 km. Thereafter, the velocity remains more or less constant uptil a depth of 23 km. Below this depth the velocity increases again reaching a value of 7.15 km/sec at 40 km. Therefore a crust comprising of three layers of thicknesses 15, 8 and 15 km was considered. As regards the heat production value in the top layer, a mean value of 4.3 HGU has been considered based on those obtained for near surface rocks (7.0 HGU) and for the base of the upper crust of stable continental areas 1.7 HGU as suggested by Allis (1979). Such aprocedure was adopted in order to account for continuous increase of compressional wave velocity upto a depth of 15 km.

The Cuddapah basin is a distinctive feature of the Indian Shield. DSS studies over the Basin have shown a basement depth of about 8-10 km. Thus three-layers crustal models (C_1 and C_2, Table IV B-3) have been considered. For the first model the mantle heat flow is assumed to be of the same order

TABLE IV B-3

MODELS FOR TEMPERATURE-DEPTH CALCULATIONS

Model Nos.	Greenstone Belts		Peninsular Gneiss			Cuddapah Basin	
	G_1	G_2*	P_1	P_2*	P_3	C_1	C_2
Surface heat flow HFU	0.6	0.6	0.31	1.31	1.31	1.8	1.8
Generation of top radioactive layer HGU	0.6	0.6	7.0	7.0	4.3	6.2	6.2
Thickness of top radioactive layer (1st layer Z_1)	11	-	11	-	15	10	10
Thermal conductivity of 1st layer mcal/cm sec°C.	6.5	6.5	6.5	6.5	6.5	6.5	6.5
Depth to Mantle, Km	38	38	38	38	38	38	38
Heat generation of 2nd layer HGU (A_1)	0.5	-	0.5	-	1.7	5.25	1.7
Thermal conductivity of 2nd layer mcal/cm. sec°C.	5.5	-	5.5	-	5.3	5.5	5.5
Thickness of 2nd layer, km	-	-	-	-	8	13	13
Heat generation of 3rd layer HGU(A_2)	-	-	-	-	0.5	0.5	0.5
Thermal conductivity of 3rd layer mcal/cm.sec°C.	-	-	-	-	5	5	5
Mantle heat flow HFU	0.41	0.41	0.41	0.41	0.46	0.41	0.9

*Exponential model   1 HFU = 10^{-6} cal/cm^2 Sec;
1 HGU = 10^{-13} cal/cm^3 Sec.

TABLE IV B-4

TEMPERATURES (°C) VERSUS DEPTH FOR THE MODELS IN TABLE IV B-3

Ref.Temp. = T°

Depth (km) Model No.	Green Stone Belts		Peninsular Gneiss			Cuddapah Basin	
	Step model 2 layers G_1	Exp. model - G_2	Step model 2 layers P_1	Exp. model - P_2	Step model 3 layers P_3	Step model 3 layers C_1	Step model 3 layers C_2
5	45	47	88	90	94	127	127
10	88	92	150	163	170	230	230
15	134	137	199	225	231	324	332
20	179	180	246	279	285	395	427
25	222	223	291	330	355	445	519
30	263	266	334	377	384	491	612
35	302	309	375	423	430	535	702
38	324	335	400	450	457	561	755

(0.41 HFU) as that obtained for the Karnataka Craton. This assumption implies high heat generation in the middle crustal layers. However, there is no evidence direct or otherwise, for such a case.

The SW margin of the basin is characterised with low heat flow values (0.64 HFU, Table IV B-1). Also, a heat flow value of 1.2 HFU has been reported from a borehole of 252 m depth near Bandalamattu (Verma et al. 1969). A third normal value has been obtained (Gupta under preparation). Temperatures have been therefore calculated, both for low as well as high heat flow values from the mantle (Tables IV B-3 and 4).

Various models can be built for example, assuming an unchanging temperature at the Moho discontinuity under various geotectonic units of the Dharwar Craton. These, however, lead incompatible results for heat source distributions. Models which assume different heat source distributions and even crustal layering, but are restricted to constant mantle contribution to surface heat flow, are found compatible for the Dharwar and Peninsular Gneiss super groups. As regards the area near the NE margin of the Cuddapah Basin, it seems likely that the mantle conducts significantly more heat flow than 0.41 HFU.

4B.6 Conclusions

A graph of surface heat flow values and heat generation values for the Dharwar Craton shows a heat flow province with an intercept (Q_o) of 0.54 HFU and a slope of 11.3 km. The results of calculations of crustal temperatures are: 1) that the Karnataka Craton is characterised with low temperatures of the order of 400 $\pm$ 50°C at the crust - mantle boundary; 2) that temperatures under Peninsular Gneisses appear to be higher than those underlying areas of Dharwar Schists belts; 3) that high crustal and Moho temperatures alongwith large lateral temperature variations prevail under the NE margin of the Cuddapah Basin.

It appears reasonable that lateral variations in subsurface temperatures near the NE margin of the Cuddapah basin cause thermal stresses, which may be partly responsible for the occurrence of shallow seismic activity near Ongole.

Lateral crustal inhomogeneties have been revealed in the Dharwar Craton by geological, gravity, and deep seismic sounding data. These inhomogeneties are the consequence of initial P&T conditions of the earth and subsequent geochemical fractionation. The geothermal data presented here also support the existence of lateral inhomogeneties.

CHAPTER IV C

SURFACE HEAT FLOW OVER PRECAMBRIAN SHIELDS

4C.1 Introduction

Our knowledge of the surface heat flow of the earth has grown significantly during the last fifteen years, the number of heat flow determinations increasing from about 1000 in 1965 to over 7,000 by the first quarter of 1980. The new results whilst making only a marginal difference to the average values of global heat flow i.e., from 63 mW/m^2 (1.5 HFU) Lee and Uyeda (1965) to 74.3 mW/m^2(1.77 HFU). Jessop et al. (1975), do however contribute valuable information in respect of the tectonic setting of a region and the thermal evolution of the earth.

With increasing number of heat flow determinations, 124 at present, the average value of heat flow over Pre-Cambrian shields has risen marginally but progressively from 34-38 mW/m^2 reported by Karaskovskiy (1961); 39 mW/m^2 by Polyak and Smiynov (1968); 40 mW/m^2 by Gupta, Gaur and Narain (1974); Rao and Jessop (1974); to 41 mW/m^2(0.98 HFU), Gupta and Gaur, 1980 (communicated and this chapter table IV C-1). However, notwithstanding this low average value indicative of the general pattern of heat flow, certain anomalous zones of high and very low heat flow do exist in the shield areas.

In order to gain a better insight into the present thermal field of the Indian Shield generally and the Aravalli tectonic belt in particular, a study was made of the surface heat flow in the Precambrian shields in the light of geological and other information of specific areas. An attempt has also been made to interpret the anomalously high heat flow values, observed in the Delhi Super Group of the Aravalli belt in terms of its evolutionary history and the attendant earth processes. These results are presented in this chapter.

4C.2 Distribution of Heat Flow in Shield Areas

From a fair number of heat flow data now available from the Indian, South American, Australian and African Shields of Gondwanaland, and the Canadian, Baltic, Ukrainian and Greenland shields of Laurasia, average heat flow values were computed for various shields and shield blocks of Gondwanaland and Laurasia (Table IV C-1). In all, 124 values were used for this computation and analysis, carefully excluding those which

i) were suspected to involve methodological errors.

ii) pertained to a zone of a local anomaly, e.g., Koshya and Khibini Mounts, Kola Peninsula USSR etc., and

iii) belonged to a region of poorly known tectonic context.

It is interesting to point out that the Gondwanaland shields show higher heat flow values than their Laurasian counterparts, the highest being 54 mW/m^2(1.29 HFU); 52 mW/m^2 (1.24 HFU) associated with the Indian and South American Coastal shields respectively (vide Table IV C-1).

Tables IV C-2 and 3 respectively give the distribution of available heat flow data separately for Archaean and Proterozoic areas in respect of the Gondwanic and Laurasian shields. These show that the shield segments of proterozoic age are associated with higher heat flow than those of the Archaean areas both in the Gondwanic as well as the Laurasian shield blocks (Tables IV C-2, 3 and 4).

The frequency distribution of the data over i) the Gondwanic shields, ii) the Laurasian shields, and iii) the two together is given in Table IV C-5. Three heat flow zones characterized by a) low heat flow values in the range of 25 to 40 mW/m^2(0.6 to 0.95 HFU), b) intermediate values in the range of 40 to about 60 mW/m^2(0.95 to 1.44 HFU), and c) high values ranging from about 55-60 mW/m^2 upwards are distinguishable. The table shows that the frequency decreases with increase in the heat flow values.

4C.3 Thermal Structure of Shield Areas

Shields are continental blocks of the earth's crust which have been relatively stable for over 600 m.y. and in contrast to the strong fk folding suffered by adjoining geosynchinal belts, have under gone only gentle warping.

TABLE IV C-1

HEAT FLOW DISTRIBUTION FOR DIFFERENT PRECAMBRIAN SHIELDS

	n	q range	$\bar{q}$	σ	s
a) Gondwanic Shields	48	18 - 75	47	1.9	13
i. Indian Shield	14	27 - 75	54	4	15
ii. South American Coastal Shield	7	37 - 69	52	4.5	--
iii. Australian Shield	13	29 - 54	40	2.3	8
iv. African Shield	14	18 - 60	44	3	12
b) Laurasian Shields	76	20 - 54	37	1	8
i. Canadian Shield	31	20 - 54	37	2	9
ii. Baltic Shield	21	21 - 50	37	2	9
iii. Ukrainian Shield	22	25 - 54	37.6	2	7
iv. Greenland Shield	2	36 - 43	39.5	4	5
c) Total values	124	18 - 75	41	1	11
d) Climatic Effect					
Canadian Shield (Cor.) SUB Prov.	20	25 - 60	40	2	8
Canadian Shield (u.cor) SUB Prov.	20	20 - 50	34	2	8

NOTE: n = the number of values

$\bar{q}$ = the arithmetic mean value

σ = the standard error of the mean

S = the standard deviation

(All heat flow values are in mW/m^2; $1\ mW/m^2 = 0.02388\ \mu cal/cm^2 sec$).

TABLE IV C-2

HEAT FLOW DISTRIBUTION FOR DIFFERENT SHIELDS OF GONDWANALAND

	n	q range	$\bar{q}$	σ	S
a) Archaean					
i. Indian Shield	5	32 - 55	43	4	8
ii. South American Coastal Shield	3	37 - 58	42	4.5	8
iii. Australian Shield	13	29 - 54	40	2.3	8
iv. African Shield	12	18 - 54	41	3	11
Total values	33	18 - 58	41	2	9
b) Proterozoic					
i. Indian Shield	9	27 -75	60	5	15
ii. South American Coastal Shield	4	48-69	59	5	9
iii. African Shield	2	59 - 60	59.5	0.5	0.7
Total values	15	27 - 75	59.5	3	12

n, q, σ, S : Please refer to Table IV C-1

TABLE IV C-3

HEAT FLOW DISTRIBUTION FOR DIFFERENT SHIELDS OF LAURASIA

	n	q range	$\bar{q}$	σ	S
a) Archaean					
i. Canadian Shield	20	20 - 50	34	2	8
ii. Baltic Shield	6	29 - 37	32	2	4
iii. Ukrainian Shield	22	25 - 54	37.6	2	7
Total values	48	20 - 54	35	1	7
b) Proterozoic					
i. Canadian Shield	11	33 - 54	42	2	8
ii. Baltic Shield	15	21 - 50	40	3	10
iii. Greenland Shield	2	36 - 43	39.5	4	5
Total values	28	21 - 54	40	2	9

n, q, σ, S : Please refer to Table IV C - 1

TABLE IV C-4

HEAT FLOW DISTRIBUTION

		n	q range	$\bar{q}$	σ	S
I	Gondwanic Shields	48	18 - 75	47	2	13
	a) Archaean	33	18 - 58	41	2	9
	b) Proterozoic	15	27 - 75	59.5	3	12
II	Laurasian Shields	76	20 - 54	37	1	8
	a) Archaean	48	20 - 54	35	1	7
	b) Proterozoic	28	21 - 54	41	2	9
III	Total values	124	18 - 75	41	1	11
IV	Climatic Effect					
	Canadian Shield Sup. Prov. (u. cor)	20	20 - 50	34	2	8
	-do-(corrected)	20	25 - 60	40	2	8

n, q, σ, S : Please refer to Table IV C-1

TABLE IV C-5

FREQUENCY DISTRIBUTION OF SURFACE HEAT FLOW DATA OF PRECAMBRIAN SHIELDS

Units : $mW\ m^{-2}$

Range No. of values	Upto 25	26-30	31-35	36-40	41-45	46-50	51-55	56-60	61-65	66-70	Over 71	To-tal
Gondwanic Shields	2	3	2	10	5	9	6	4	3	1	3	48
Laurasian Shields	7	7	21	12	17	8	4	-	-	-	-	76
Total values	9	10	23	22	22	17	10	4	3	1	3	124

The observed variation of heat flux over shield areas can be attributed mainly to a) their initial growth pattern, b) initial degree of partial differentiation and migration of the heat producing elements K, U and Th from deeper levels and their distribution in the crust and mantle prior to 600 m.y. and c) uplift and erosion, which have been continually active since their very formation. Although various hypotheses have been advanced from time to time to explain the formation of continents, not much is definitively known regarding their growth patterns. However, recent spurt in geochemical investigations of shield areas, owing to their recognition as potential repositories of mineral wealth, have shed significant light on the nucleation of continental masses. It is now believed that a central mantle plume or a family of plumes provided the nuclei for continental masses which subsequently accreted and merged into proto-continents Goodwin (1968, 1971), Pavlovskiy (1970), Anhaeusser (1971, 1973), Glikson (1970, 1971). The Precambrian crust of the earth is thus envisaged as having evolved through six super shields or proto-continents which were, in turn, formed through concentric growth with some parts experiencing orogenic events and geosynclinal phases. The Precambrian crust was, perhaps, the first to become continuous throughout the super continents of Laurasia and Gondwanaland (Dietz and Holden, 1970; Goodwin, 1974). These two must have been then united to form a single continental mass of Pangaea, envisaged by Wegener (1912) to be subsequently split and set adrift towards their present positions.

Heier (1965), Lambert and Heier (1968) showed, on geochemical evidence, that heat producing radioactive isotopes were concentrated upwards in the continental crust and Birch et al. (1968) demonstrated that a linear relation exists between the value of surface heat flow and the heat production caused by the decay of radioactive isotopes U^{238}, U^{235}, Th^{232} and K^{40} in the near surface rocks. A consistent relationship between continental heat flow and the age of the last tectonothermal event was also shown to hold, by Lee and Uyeda (1965), Gupta (1967), Polyak and Smirnov, (1968), Hamza and Verma (1969). A number of studies have appeared in recent years analysing the surface heat flow in terms of the likely source distribution in different depth sections of the earth, Crough and Thompson (1976), Kutas (1977), Kono and Amano (1978), Vitorello and Pollack (1980).

Polyak and Smirnov (1968) on the basis of 446 continental heat flow values, showed that tectonic provinces of different ages are characterised by different average values of heat flow, being as high as 93 mW/m^2 for younger structure such as the Cenozoic Volcanic provinces and as low as 39 mW/m^2 in regions of Precambrian folding. Subsequently numerous writers ascribed this variation to two processes, viz., i) erosion and removal of the top radioactive layers, and ii) the decay of thermal perturbation introduced by the tectonicity of the province. Vitorello and Pollack (1980) demonstrated that this decay process could be reasonably represented by a time scale of $q(t) = a.t^{-\frac{1}{2}}$ as happens to hold good over the oceanic areas. However, the two processes combined have no simple time scale representation. It should ofcourse be obvious that variations in surface heat flow over Precambrian regions have nothing to do with tectonically induced thermal perturbations and arise from variations in the amount of crustal radiogenic heat production and the deep heat rising from below the zone of crustal radioactive enrichment.

Various questions therefore arise while attempting to explain the variations in heat flow values observed in different shield regions, i) was the character of the initial mantle plumes uniform? ii) Did significant differences in the chemical composition of continental masses occur at the time of initial cratonisation of the shields? iii) Is the mantle component of heat flow different under f different shields? Or can the variation in amounts of concentration of radioactive elements in the near surface rocks explain these differences? iv) Do the observed high heat flow values at certain locations in the shields reflect some characteristic geological environment and if so what are the significant geological parameters? Definitive answers to these questions are clearly not available today but most likely, some combination of these processes occurred in cohort.

Different areas of similar thermal regimes as well as dissimilar ones, notably the Gondwanic and Laurasian shield areas have been compared and studied to elucidate some of these questions.

4C.3.1 Areas of low heat flow

Such areas are mostly occupied by reworked Archaean or Archaean, basic and ultrabasic metamorphic rocks, and greenstone belts. Salop and Scheinmann (1969) state that whilst rocks of the amphibolite and granulite facies which have lower concentration of Uranium and Thorium are well developed in the Archaean-complexes, these are absent or only locally developed in post-Archaean and Precambrian deposits. Uranium Thorium and Potassium which were more progressively concentrated in the uppermost portions of the crust by upward migration during the Precambrian, were continually depleted as erosion progressed apace. For example, the Canadian shield whose cratonization according to Goodwin (1974), was substantially advanced by the end of Archaean (2,500 m.y) and completed by the end of the Aphelian time (1,735 m.y), is characterised by a low value of mean heat flow. Edwards and Hasan (1970) have indicated that 90% of all rocks exposed in the structural provinces of the Canadian Precambrians were dated to be Archaean but were reworked at different times. By using structural, textural and geochronological technique, Burwash et al. (1973) have demonstrated that broad areas of the Churchill Province of the Canadian shield have been remobilised. Quantitative estimates by Shaw (1967) of the mean concentration of heat producing elements in the Canadian shield also show that its surface rocks are characterised by a low value of heat production equal to 1.63×10^{-6} mW/m^3.

The stages of evolution of the greenstone belts which are mainly occupied by metavolcanic basic and ultrabasic rocks associated with some sedimentary groups and granulite facies have been entirely different, their association with low heat flow being mainly a chance coincidence.

The lowest reported heat flow values so far in the Precambrian shields are:

i) 18 and 21 mW/m^2 in the West African Craton, Cahpman and Pollack (1974),

ii) 21 mW/m^2 in the Baltic shield by Swanberg et al. (1974).

The latter has been observed in boreholes drilled into and around a titanium bearing ilmenite orebody located near the central part of a 1,000 km^2 anorthosite complex of deep seated origin. The correction for paleoclimatic temperature variations is less than 0.4 mW/m^2. Various possible phenomenon including impoverishment of lower-crustal and upper mantle heat generating materials have been suggested for this very low heat flow value. A low heat generation value of 0.09×10^{-6} Wm^{-3} was obtained by Swanberg et al. (1974) for the rock types of the area, which is similar to that measured in anorthosite rocks from Holenarasipur, Dharwar schist belt (Table IV B-2)'

4C.3.2 Areas of intermediate heat flow

Intermediate heat flow values in shield areas of the order of 40-60 mW/m^2, have been generally observed in two distinct geological environments. The first of these are related to bodies of granites or granitic gneisses or to granitic plutons occurring at shallow depths. This is quite understandable as granitic rocks are more radioactive (heat generation being 2.6×10^{-6} Wm^{-3}, Heier and Rogers, (1973) and more enduring owing to their resistance to weathering and erásion.

Intermediate heat flow values have also been observed in the marginal regions of shields quite irrespective of the nature of rock formations. A value of 55 mW/m^2 has been reported by Beck and Sass (1966) over the Muskox intrusion, in the Canadian shield, consisting of mafic and ultramafic rocks. This, as well as slightly lower values have been observed in the marginal regions of the Western Canadian shield (Lewis, 1969; Jessop, 1968), whilst the central part of the shield (Superior Province) is characterised by average heat flow of the order of 40 mW/m^2. This anomaly is caused by the fact that the margins of the shield blocks which are bounded by faults experienced recent thermo-tectonic events.

4C.3.3 Areas of high heat flow

The main areas of highm heat flow (70 mW/m^2) have been found to occur in two regions. These are in the Indian shield, the Delhi system (this thesis, section 4A.6.1) and

the north

the northeastern margin of the Cuddapah basin. The Delhi system consists of folded rocks of Upper Proterozoic age which were deposited in a geosyncline formed in a newly collapsed area within an already complex Archaean Lower Proterozóic basement. Thus, at the time when other portions of the shields were undergoing processes of metamorphism, uplift, and erosion, some parts of the shields such as the Delhi belt were sinking and receiving radioactive sediments eroded away from uplifted portions, some of these strata being preserved even to this day. According to Sarkar (1968, 72) the Delhi orogenic cycle closed at around 750 m.y and the Khetri phase, region for which heat flow values were determined slightly later (Ca 600 m.y.). The Delhis which are about 6 km thick (Pichamuthu, 1967) with a mean heat generation capacity equal to $2.2 \times 10^{-6} Wm^{-3}$ (Rao et al. 1976), contribute about 13 mW/m^2 (0.31 HFU) towards the observed heat flow. Considering heat flow and heat generation data from regions of diversified geology and tectonic setting, widely separated from each other, Rao et al. (1976), estimate that the thickness of the top radioactive layer for these regions including the Khetri area to be ~~15~~ 14.8 km. There is no geological evidence that the Delhis are so thick in the Khetri copper belt, being exposed in a long narrow belt extending approximately NE-SW from near Delhi to Gujrat and overlying the Aravalli ~~s~~ or Pre-Aravalli Systems (Sections 4A.2 and 4A.3) exposed in their vicinity. Measurements in these regions have yielded normal heat flow values equal to 41 mW/m^2 (sections 4A.6.2 and 4A.6.3) indicating thereby that much of the top radioactive layer of the exposed Aravalli and Pre-Aravalli systems have probably been eroded. It may be that the underlying layers contained more radioactive elements in comparision with their counterparts exposed on the surface, but their original concentration must have changed due to later processes of metamorphism, melting and magmatism. It appears probable that under such portions of the shield which experienced geosynclinal developments during the middle Proterozoic, the lithosphere has not been differentiated as much as it has been under other shield areas, thereby causing the mantle component of heat flow to vary in different regions of the Aravalli Precambrian Craton, and the isothermal interface between the lithosphere and asthenosphere to lie at different depths.

The Cuddapah basin which is an intra-cratonic Proterozoic basin, also depicts a very contrasting geothermal character, where a heat flow value of 75 mW/m^2 intermediate to low, was obtained. The estimate of 14.8 km as the thickness of the top radioactive layer for the Cuddapah basin would be incompatible on geological and geophysical grounds, as the thickness of the Cuddapah sediments, revealed by deep seismic sounding studies, varies from about 8 to 10 km. (Kaila et al. 1979). An evolutionary history similar to that of the Delhis, together with granitisation of the Cuddapah Sediments, which are analogous to the Delhis, appears to be the cause of high heat flow (75 mW/m^2) at the northeastern margin of the basin.

4C.4 Comparison of Heat Flow Between Gondwanic and Laurasian Shields

It has already been pointed out that the Laurasion shields are generally associated with lower heat flow than the Gondwanic shields (Table IV C-1 and 4). As the former lie nearer the pole, it is probable that these have suffered greater uplift and subsequent erosion following deglaciation, or that the corrections to be applied in respect of past climatic changes, although not used in all cases, are not adequate and need to be reconsidered. It is worth noting that Jessop and Lewis (1978)(Table IV C-6) demonstrated an 18% increase in the magnitude of the measured heat flow value if past climatic effects were taken into account. If it were so, corrections of the mean heat flow values in the Laurasian shields should not be statistically different from those for the Gondwanic shields. However, observations in a 3 km drill hole revealed no vertical variation in the heat flow (Sass et al. 1971) which may be attributable to past climatic changes.

Whilst uneven sampling of heat flow data sets, can not be ruled out as being the cause of high mean heat flow obtained over some of the shields notebly, the Indian and South American coastal shield, there are other Precambrian regions showing significantly different thermal characteristics. The African land mass consists of three large Precambrian Cratons viz., the Kalahari, The Cango and The West African Craton. Each is roughly equidimensional and of the same size as the

TABLE IV C-6

PARAMETERS OF HEAT FLOW PROVINCES

Province	n	Q_o (mWm^{-2})	b (km)	Reference
Shields of Gondwanaland				
i. Indian Shield				
a) Dharwar Craton	4	23	11.3	Gupta & Sharma(1979)
b) Proterozoic shield (?)	6	39	14.8	Rao et al.(1976)
ii. * South American Coastal Shield	4	28	13.1	Vitorello, Hamza, Pollack (1980)
iii. Australian Shield				
a) Western Shield	4	26	4.5	Jaeger (1970)
b) Central Shield	10	27	11.1	Sass et al.(1976)
Shields of Laurasia				
i. Canadian Shield				
a) Superior Province	3	21	13.9	Cermak and Jessop (1971)
b) -do-	11	21	14.4	Jessop & Lewis(1978)
c) -do-(corrected for climatic changes)	11	28	13.6	-do-
ii. Baltic Shield (Norway)	4	20	8.4	Heier and Gronlie (1977)
iii. Ukrainian Shield	6	25	8.0	Kutas (1979)

Baltic Shield. Out of 14 heat flow values, which have been considered for the African Shield, 12 are from the Kalahari Craton and the other two from the West African Craton. It is worthy of note that amongst the 29 recognised Precambrian Cratons of the earth, the Kalahari Craton is characterised with the highest heat flow (48 mW/m^2) while the reported heat flow value (18 and 21 mW/m^2) for the West African Craton by Chapman and Pollack (1974) are the lowest. Such contrasting thermal character will have to be traced back to the various stages of evolution of different segments of the African shield. The answer to this anomaly perhaps also lies in the compositional variations of lithospheric plates bearing different shileds. The close correspondence between higher heat flow over Indian, African (Kalahari Craton) and South American coastal shields would thus appear self explanatory in view of their predrift continuity.

4C.5 Reduced Heat Flow

It has been mentioned in section IV B-3 that the linear relationship between surface heat flow Q and heat generation Ao in near surface rocks has led to the division of surface heat flow into two components: one arising from a top radioactive layer and the other 'Q_0' (the reduced heat flow) from the heat flowing upward from the lower crust and mantle. The relation also led to the recognition of large geothermal provinces with constant value of 'reduced heat flow' and equal thickness 'b' of the top radioactive layer but with

1. That heat flow values over shield areas vary from 18 to 75 mWm^{-2} and are not uniformly low as was generally believed. Three characteristic regions of low, intermediate and high values can be identified.

2. That those regions of the shields, which are covered by high grade metamorphic rocks or by Archaean or reworked Archaean rocks or greenstones, are generally characterised by low heat flow values of about 20 to 40 mWm^{-2}. In the case of greenstone belts, it is the nature of their initial evolution which is mainly responsible for their low heat flow associations whereas in the Archaean high grade terrains, it has been caused by the vertical seggregation of radioactive elements and their subsequent depletion with erosion of overlying layers. According to Watson (1978) the magnitude of the Post-Archaean vertical movements required to bring the Archaean granulite complexes to their present eroded level, suggests that their subsequent evolution followed a different course from that of the greenstone belt provinces which have maintained a rather constant level. Another pointer to the operation of different thermal regimes during the evolution of various segments of the Precambrian crust is furnished by the occurrence of diamonds, which reach the surface in regions of low thermal gradients. Considering the stability field of diamond occurrences another writer, Shackleton (1973) suggests that despite a generally high thermal gradient within the shields, there existed some regions where it was low during the early Proterozoic times.

3. That the Proterozoic regions of the shields are characterised with higher heat flow than the Archaean. Apart from other considerations, it can be mentioned here that on the basis of detailed geochemical studies, numerous workers have pointed out that the earlier crust has been more basic, had less K, U and Th and more Na, Cr, Ni and perhaps Au, than the younger crust.

4. That granitic regions of the Precambrian shields are associated with higher than average heat flow (1.1 to 1.8 HFU) due, mainly, to the concentration of high radioactive elements in the granitic gneissic rocks.

Furthermore, marginal regions of shields, Canadian and Ukrainian for example, are associated with higher heat flow than their central regions lending credence to the suggestion of concentric growth of continental crust. Besides, heat flow values as high as 55 (1.31 HFU) have been observed even in the basic rocks located near the margins of the shields due perhaps to subsequent thermo-tectonic processes occurring along marginal faults.

5. That Proterozoic miogeosynclinal belts, Indian shield for example, have been found to mark the highest heat flow regions in the shields. It might be that in such regions, erosion of the top layers of the sediments, which were deposited and metamorphosed in the upper Proterozoic, was not severe enough to deplete their radioactive contents appreciabley. However,

it appears more likely to be the result of the lithosphere being less differentiated under such portions of the shield which experienced geosynclinal developments during the Middle Proterozoic.

6. That the 'reduced heat flow' and the mantle component of heat flow is likely to be different under different shields and or parts thereof. This implies that the isothermal interface, between the lithosphere and asthenosphere may be an undulating surface.

7. That the Gondwanaland shields have in general, a higher heat flow than their Laurasian counterparts. However, the difference between the heat flow in the two shields blocks becomes insignificant if climatic corrections are applied to all values obtained over the Laurasian shields. The Kalahari Craton of the South African shield is associated with 40% higher heat flow than the Baltic and the Ukrainian shields.

CHAPTER V

GEOTHERMAL RESOURCES OF INDIA

5.1 Introduction

India is a big country with a large population, where local conditions and availability of natural resources vary greatly from one region to another, particularly of Energy resources which constitute a crucial input for development. The only way to provide it to every corner of the country, whether in the inhospitable desert regions or in the remotest regions of perennial snows, without wastage and without degrading the environment is to develop and exploit widely distributed sources such as solar and geothermal energy, wind and tidal power and biogas. The relevance of our R&D efforts in these fields thus acquire a special significance.

Although geoheat has been turning the blades of turbines and generating cheap power since quite some time in Italy, New Zealand, USA, USSR and Japan, it did not attract much attention in other parts of the world until recently, when the implications of the new global energy crisis burst upon us with a bang.

Geothermal exploration seeks to delineate regions at shallow depths where thermal energy is concentrated. An examination of the well explored and developed geothermal fields of the world indicate that most of these are located in young orogenic zones associated with Quaternary volcanism. However, Larderellow, the first well-explored geothermal field in the world, lies at a considerable distance from the nearest volcano. This area and a few others have been found in the hinterland fault block structures or hinterland basins and in rift zones. Potential geothermal areas are also known to occur in regions of crustal rifting and of recent mountain building. Most of the geomechanical energy released in the earth is expended within a few narrow orogenic belts marked by a strong seismic activity. These are the junctions of large crustal plates and are highly deformed along their borders. In their interior regions, mostly broad epirogenic movements occur. The heat energy is concentrated mostly at the plate margins, or at places within the plates marked by zones of ancient weakness or of magmatic activity.

No recent volcanic activity is known in India except in the Barren Islands which form the crest of a deeply submerged ridge in the Bay of Bengal. However, many hot springs have been known in this country from ancient times and plate boundaries, rift valleys and tectonic zones affected by recent crustal movements are also found to occur in several regions. All these are generally associated with hydrothermal systems and point to the potential and possibilities of an exploration, evaluation and assessment programme of geothermal resources.

In order to gain a better insight into the geothermal resources of India and to recognise, if possible, certain geothermal provinces of India, where subsurface thermal waters may be encountered and exploited, an analysis and synthesis has been attempted of available geodata, particularly of the major tectonic features of the Indian land mass, of the surface manifestations of geoheat and of available heat flow data in the Indian peninsula. From these, it is felt that nonvolcanic geothermal resources of moderate to low grade can be expected to occur in certain geothermal provinces of India.

5.2 Evolution of The Indian Landmass

The Indian landmass has been divided both geologically and geographically into the following three main provinces.

1. Peninsular India
2. The Indo-Gangetic alluvial plain } Extra Peninsular India
3. The Himalaya }

Potential regions of geothermal resources are those which continued to be tectonically active during the Tertiary-Quaternary periods and or suffered x Pliocene - Holocene igneous and volcanic activity. Accordingly, the evolution and growth of the above mentioned geological units of Indian landmass have been summarised below. Tectonic events, wherever relevant to creation of geothermal resources, have also been discussed.

5.2.1 Peninsular India

The creation of continental masses and the dyanmism of the earth as a whole have been primarily controlled by its internal heat. It is believed that deep-seated, long lived convective mantle cells, and mantle plumes produced continental nucleii and protocontinents, which merged at one time to form a single continental mass, the Pangaea. At the end of Triassic, the Pangaea had split into Laurasia and Gondwanaland (Dietz and Holden, 1970). The latter itself broke during the early Cretaceous. According to McElhinny, (1973), Klootwik (1973) and Valencio (1975), the Gondwanaland appears to have remained mostly intact from the late Pre-Cambrian into the Mesozoic, although the first indications of the eventful break-up may have appeared in the late Paleozoic. Craddock (1977) described twelve events in the Gondwanaland fragmentation, which possibly first occurred with the separation of west Gondwanaland (South America - Africa) from East Gondwanaland (Antarctica-India-Australia). The time of separation of India from Antarctica is difficult to establish but according to Craddock (1977) it probably occurred in early Cretaceous. McElhinny (1976), puts the event around 130 m.b.B.P., while Laughton and others (1973) date this separation to be older than 75 m.y.B.P.

The break-up of the Gondwanaland was attributed to deep thermal processes. Hot spots have been believed to be the cause of rifting between Australia, India and the

Antarctica. According to Morgan (1972) Gondwanaland may have fractured along rifts that made triple junctions over rising mantle plumes. Numerous hot spots are believed to be still operative in the earth's mantle under various places.

As regards the Indian Shield, six continental nucleii are believed to have developed during the early stages ($>$ 3.5 billion years) of its crustal evolution. These can be grouped into three pairs, demarcated by the Narmada-Sone and Godavari linements. Later, through vertical and horizontal accretion these three pairs developed into three protocontinents, the Dharwar, the Aravallis and the Singhbhum, Raja Rao (1971) and Naqvi, Rao and Narain (1974).

Regarding the nature of the primordial crust of the Indian shield, various suggestions have been put forward. One group favours that the segments of the Indian proto-crust may be anorthosites, or basic and ultrabasic charnokites of komatiitic chemistry, while others consider it to be made up of granities, gneisses, migmatites, oceanic tholeiites, basaltic komatiites, Detailed geochemical work at the NGRI supports the view that the primitive crust was thin, unstable, basaltic and of oceanic type; that the Archaean mantle beneath southern Indian was peridotitic; that late Archaean to lower Proterozoic was transitional from simatic to sialic crustal development; and that Cratonisation was over by mid-Proterozoic (Pichamuthu 1977).

Krishnan (1953) and other earlier workers delineated four persistant regional trends in the Archaean - Proterozoic rocks of India. These are:

1. The Satpura trend (ENE-WSW) in central and eastern India extending from Panchmahals (Gujrat) to the Assam Plateau.

2. The Eastern Ghats trend (NE-SW) along the Eastern Ghats of Orissa, Andhra P and Eastern Madras.

3. The Aravalli trend (NE-SW) in Rajasthan.

4. The Dharwar trend (NNW-SSE) in Mysore and Hyderabad, extending as NW-SE trend into southern Madras and Kerala.

It is believed that two or more regional trends intersect or intermingle at certain locations notebly in areas like the Bhandara triangle, the Mahanadi valley, the Nellore mica belt, southern Mysore and Wynad.

Eremenko and Negi (1968) view the tectonic evolution of the Earth's crust in India along two main stages:

i) geosynclinal and ii) platform.

They suggest that the five Archaean-Proterozoic folded regions, which constitute the basement of Indian platform are areas of i) Dharwar folding; ii) Aravalli folding; iii) Eastern Ghat folding; iv) Satpura folding and v) Delhi folding. Areas of Dharwar orogeny have been considered the oldest followed by Aravalli, Eastern Ghat, Satpura and Delhi orogenies.

The Cuddapah sequence represent the oldest sedimentary sequence of the Indian platform and occupies the Cuddapah depression, the Godavari graben and probably the major part of the Deccan synoline and the Western part of Narmada-Sone-Damodar graben. Subsequently, positive movements resulted in complete erosion of the Cuddapah sediments from some uplifted parts of the intervening platform areas. The next stages of the tectonic evolution of the Indian platform followed through the Vindhyan (Upper Proterozoic - Lower Palaeozoic), Gondwana (Upper Carboniferous to Lower Cretaceous), the Middle and Upper Mesozoic and other sequences.

Granitic intrusive activity continued with breaks throughout the Proterozoic Era. Numerous large and small granitic bodies are exposed presently and many have been inferred on the basis of relevant geodata. The main large granitic bodies are the Bundelkhand, Erinpura, Berach, Singhbhum, Dongargarh and Closepet massifs.

A significant event in the evolution of the Indian platform is the widespread Upper Cretaceous - Lower Eocene Igneous activity, represented by the Deccan Trap lava flows, which presently cover a very large part of the Western and Southern India. Igneous activity occurred during Upper Jurassic to Lower Cretaceous also and is represented by

Rajmahal and Sylhet traps and intrusive dykes and sils in the Gondwanas. However some of these are considered contemperaneous with the Deccan Traps.

Earlier views consider the Indian Peninsula to have remained a stable shield unaffected by movements of any kind and all the physiographic peculiarities, an end result of f circum-denudation of an old table land. However, recent geoevidence indicates that eventful vertical movements occurred in the Peninsula during all times including the Tertiary and Quaternary periods. Uplift and rejuvenation of the Peninsular Shield did occur in Cretaceous and Mid-Cenezoic. Post-Miocene uplifts are noticeable in the Cuddapah basin, the Western Ghats and in the Mysore Plateau.

Sedimentation occurred in Kathiwar and Kutch area during Tertiary period but these were raised up into dry land in the latter part of Pleistocene. Uplift of the Western Ghats during the Upper Tertiary is indicated by the youthful behaviour of the upper reaches of the river basins of the area.

A major regional upwarp; whereby the older peneplanned surface was uplifted to a height of over 3,000 ft, occurred during the Mio-Cainozoic in the Peninsular shield. Laterite, which had formed a crust over the peneplanned surface, came to occupy elevations of 3,000 ft. over the Mysore plateau (Radha Krishna, 1976). A general upward tilt of the Peninsula in the south k also occurred there.

5.2.2 Extra-Peninsular India

India after its detachment from the Gondwanaland moved northward most likely during early Cretaceous. Laughton McKenzie and Sclater (1972) suggest that the northward movement of India started slowly but accelerated between 75 and 70 m.y.B.P. to a total rate of 17 cm/year. It decelerated and considerably slowed down quite suddenly around 56 m.y.B.P. perhaps at the time when it arrived at the subduction zone in the Tethys Sea in the north. The total drift is about 5,000 km with an average speed of about 7 cm/year. Palaeomagnetic data obtained from Indian rocks of various locations and ages (Duetseh et al. 1956; Verma, 1973) also point out to its rotation by as much as 20°. It is believed that during its northward movement in the late Cretaceous and early Tertiary, transform faults of the Owan Fracture Zone, the Mascarene-Chagos-Laccadive Ridge and the Ninety east Ridge were formed. These fracture systems extend to the great Himalayan border thrusts of Kirthar in the west and Patkai-Arakan-Yoma in the east (Gansser, 1965). The Seychells and the Chagos-Laccadive rises appear to be remnants left behind by the advancing peninsula.

The roughly northward moving Indian Shield eventually collided with the mainland of Asia, but it is difficult to establish the exact time of this event. Most workers advocate collision during the Eocene or early Oligocene (Ravi Verman, 1972; Golchen, 1975; Molnar and Tapponier, 1975; Blow and Hamilton, 1975). The origin of Siwaliks derived from the rising Himalayas, however, suggest that collision was certainly underway by the Miocene (Gansser, 1973; Verma, 1973).

The aforesaid northward movement of the Indian subcontinent after its detachment from the Gondwanaland and collision with the Eurasion plate with consequent underthrusting has played a most significant role in its subsequent tectonic history. Lofty mountain chains sprang up along the advancing edge in the north, while the sedimentary pile (Tethys) accumulated in front of the Asian craton was compressed and lined by an immense volume of ophiolites. A collateral of the mountain building activity along the northern margin of the Indian shield was the formation of a belt of persistent subsidence between the central part of the shield and the rising mountain. The Siwaliks, since raised were deposited in this depression followed by the Indo-Gangetic and the Brahmaputra basin. It is now well established that the northern margin of the old Gondwanaland was also involved in mountain building. The Himalaya in fact consists of two parts, one belonging to Peninsular India and represented presently by most of the Lesser and some parts of the Higher Himalaya, and the other belonging to the Tethys geosyncline which now occupies the northern flank of the mountain chain. Presently the Himalaya comprises of four contrasted lithotectonic units each characterised by its own distinctive structure, lithological composition and stratigraphic setting.

As a result of prolonged northern movement and collision large compressional forces at the corners of the subcontinent produced the dramatic syntaxial bends notebly the one in

Kashmir. Anatexis also occurred in certain parts of the Himalaya producing mountains such as the Nanga Parbat. The Mikir Hills - Shillong Plateau and the Ranchi Plateau were uplifted. Fusion took place in the crust under the Tethys and Higher Himalaya, and synorogenic and postorogenic granites were emplaced.

5.2.3 Geotectonic evolution and relationship with geothermal resources

It has already been mentioned that the Indian main landmass is devoid of any recent volcanism. The remarkable features of the Tertiary period of the Indian landmass have been: a) The eruption of stupendous quantities of basic lava in the Eocene which might well have begun during Cretaceous, b) compression, thrusting and uplift of the Tethys geosyncline to form a great mountain system on the northern border in the Upper Tertiary, and c) down faulting of a narrow strip along the western edge of the shield in the late Pliocene. All these must have caused considerable transfer of material in the lower crust and mantle which perhaps resulted in vertical movements, reactivation of ancient faults, and refashoning of subsurface thermal regimes.

It is now well established that the collision of Indian plate with the Eurasian plate reactivated certain ancient faults, and transcurrent tear faults such as those running through Moradabad, Lucknow, Patna and Dubri, produced block uplifts and epeirogenic movements. Imprints of this movement are infact discernible throughout the Indian shield. As believed by some, this also perhaps resulted in the outpouring of the Deccan lava flows. According to Sen Gupta and Khattri (1973), beginning in the Cretaceous, the Cambay area in Western India experienced tensional forces oriented in an east-northeast west-southwest directions, which gave rise to the formation to the Cambay rift. Ancient deep features such as the Narmada-Sone lineament continued to be reactivated. These are probably regions in which concentration of heat energy has occurred and may still be occurring at shallow crustal levels.

Deep seismic sounding studies have revealed a large crustal thicknessof about 30 km thick granitic layer under the Himalaya. Release of stresses accumulated due to great compressive forces result in the liberation of heat at the expense of tectonic friction. Northward movement of the Indian Plate caused compression, buckling, squeezing and upward movements of the Himalayan rocks during the Tertiary orogeny. Processes of compression and squeezing should have caused crustal melting resulting in generation of magma and in the emplacement of syntectonic granites under the Tethys and Higher Himalayas. Upwarping and faulting of the highly heated rocks relieved the pressure, lower in the crust, allowing fusion and generation of magma involving the depressed sialic layers. Post-Orogenic granites, notably of Badrinath-Kedarnath, Gangotri, Kalapani, Burzil, Pir Panjal, Melakarchung and Chemolhari etc., appear to owe their origin to such a process in the Himalayas.

Most of these granites seem to be of Mio-Pliocene age.

Profuse shallow seismic activity continue to occur and movements of the orogenic phase are still continuing in the Himalaya, where crustal velocities have been found to be lower compared to those in the Indian shield, particularly those under the Higher Himalay and Tethys zones, suggesting the prevalance of comparatively higher crustal temperatures under these zones. Possibilities of occurrance of Holocene zones of melting at shallow depths due to release of pressure on account of uplift and by seismic stresses in the heated crustal layers thus become fairly distinct, opening up, in turn, the possibility of finding suitable environments for the occurrence of geothermal resources, both in Peninsular India and in the Himalaya.

5.3 Thermal Manifestations in India

Surface manifestations of geoheat in the form of hot springs occur in almost all tectonic units of the sub-continent (Figure V-1). However, except for the Barren Islands, they are all devoid of any recent volcanism. Oldham first compiled and catalogued almost all the know hot springs of this sub-continent as far back as 1882. About 250 thermal springs some of which exhibit boiling with steam separation and mild geyser activity are presently known. Most of these are not isolated phenomena but follow certain tectonic trends and tend to occur in areas associated with tectonic movements.

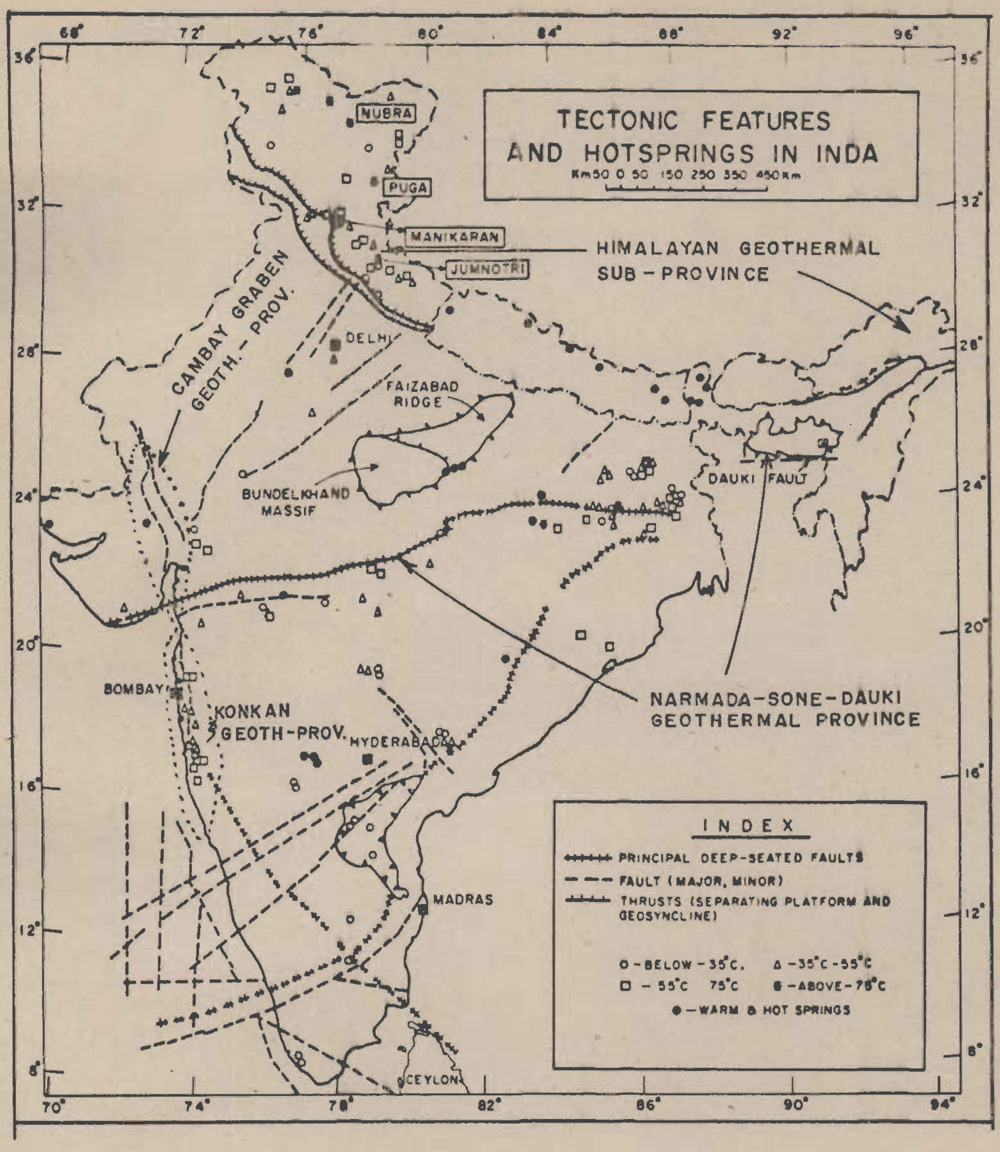


FIG. V-1 Locations of major tectonic features and hot springs in India

This is specially valid for those lying in the Himalayan Orogenic belt, in the Narmada-Sone and Damodar valleys as well as on the West Coast, commonly known as the Konkan Coast region.

Hot springs generally occurring in clusters and groups are known over 50 locations presently in the Himalayas. These are associated with thrusts and faults in its four contrasted lithotectonic units (Figure V-2). On the West Coast, many hot springs approximately aligned in a N-S direction, emerge from the shear and fracture zones near the junction of lava flows or contacts of dolerite dykes with basalts (Figure V-3). Few hot springs are also located in the Narmada and Sone valleys which mark a prominent lineament on the Indian land mass. A group of hot springs generally known as the Rajgir-Monghyr belt of hot springs, whose waters are uniquely characterised by a low pH value and low mineral content, flow from the Pre-Cambrian meta-sediments in the northeastern part of India, where crustal movements corresponding to the late phase of the Himalayan Orogeny have taken place. Another group of hot springs occur south of this belt, in an area largely covered by Pre-Cambrian crystalline rocks and partly by Gondwanas (Upper Palaeozoic - Lower Mesozoic). These include the famous Tatapani hot springs (Figure V-4), whose waters have the highest temperature (91°C) of all the springs in peninsular India. Crustal movements corresponding to the late phase/s of movements of the Himalayan orogeny have, most probably, occurred in this region. A few hot springs are also aligned along a Pre-Himalayan lineament towards south-west

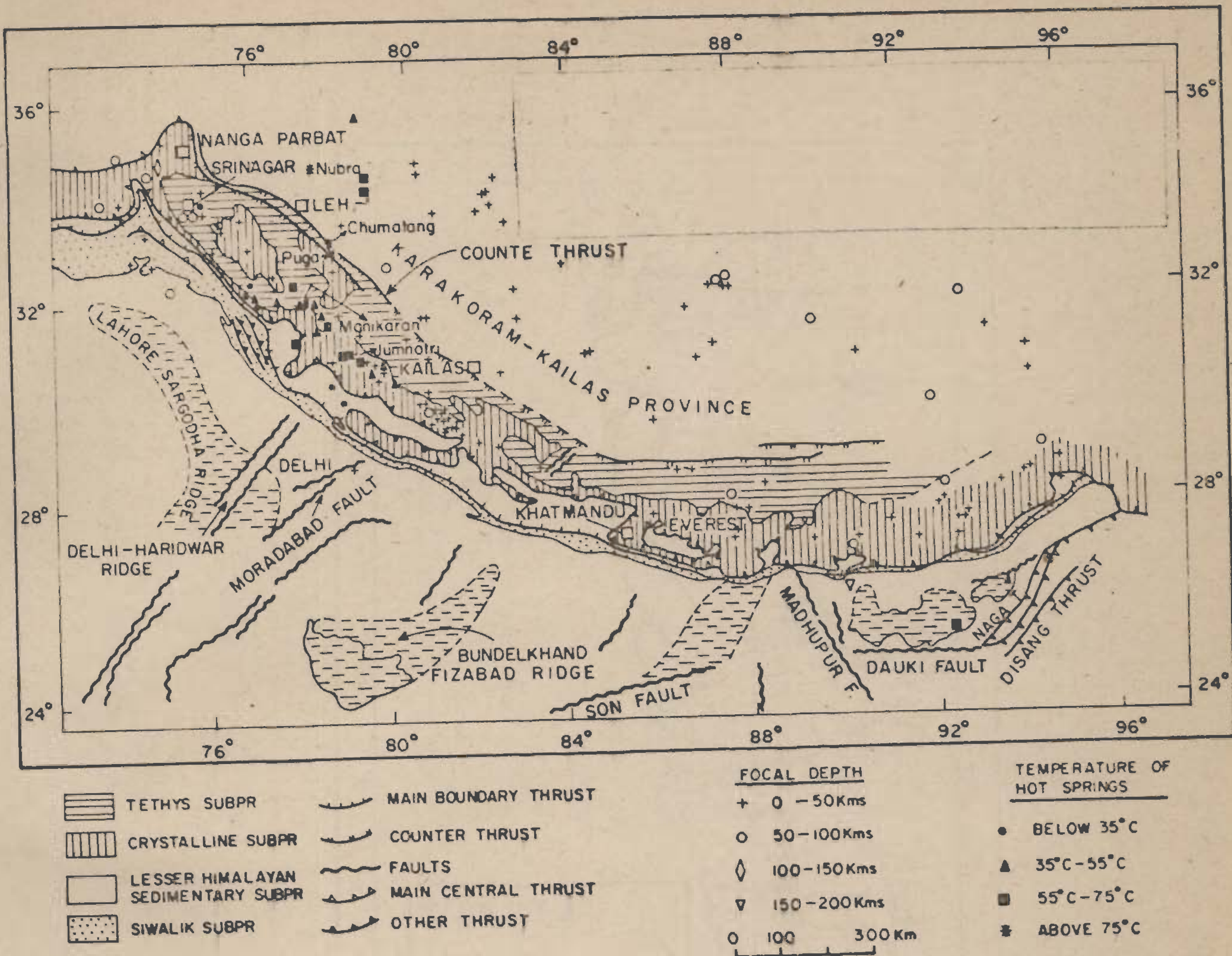


FIG. V-2 Hot springs and earthquake epicenters in the Himalayan geothermal subprovince

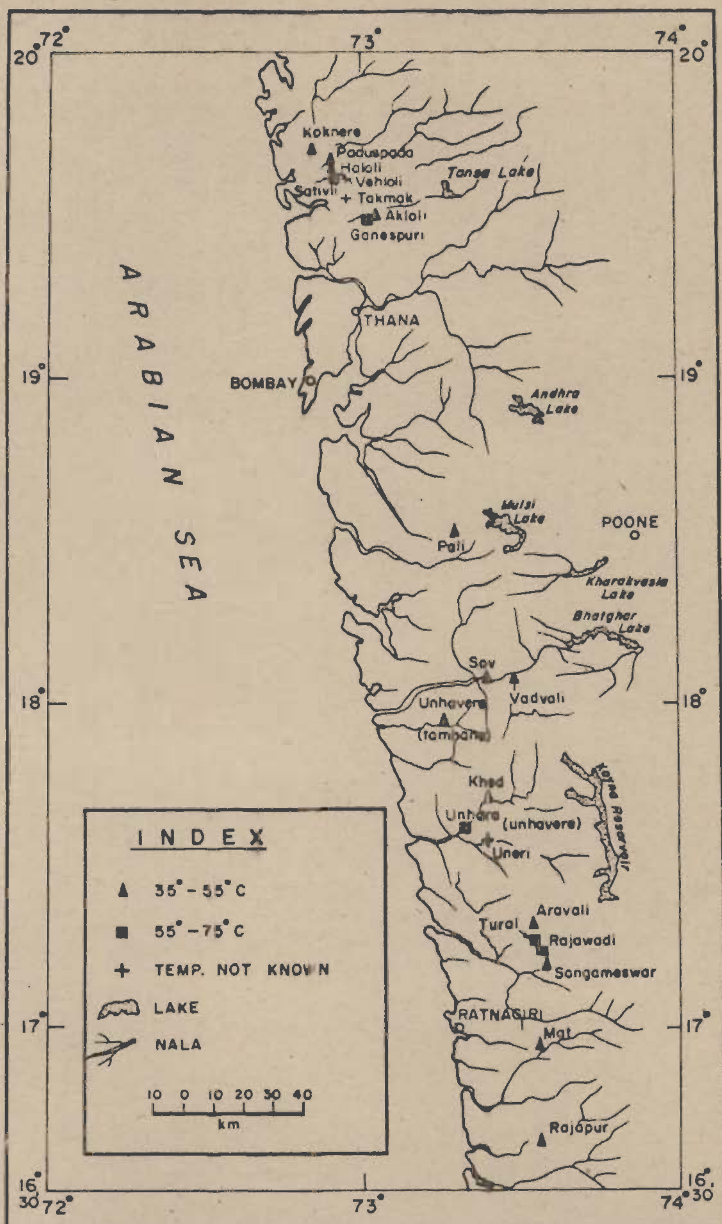


Fig. 3 LOCATION MAP SHOWING THE HOT SPRINGS IN THE WEST COAST OF INDIA.

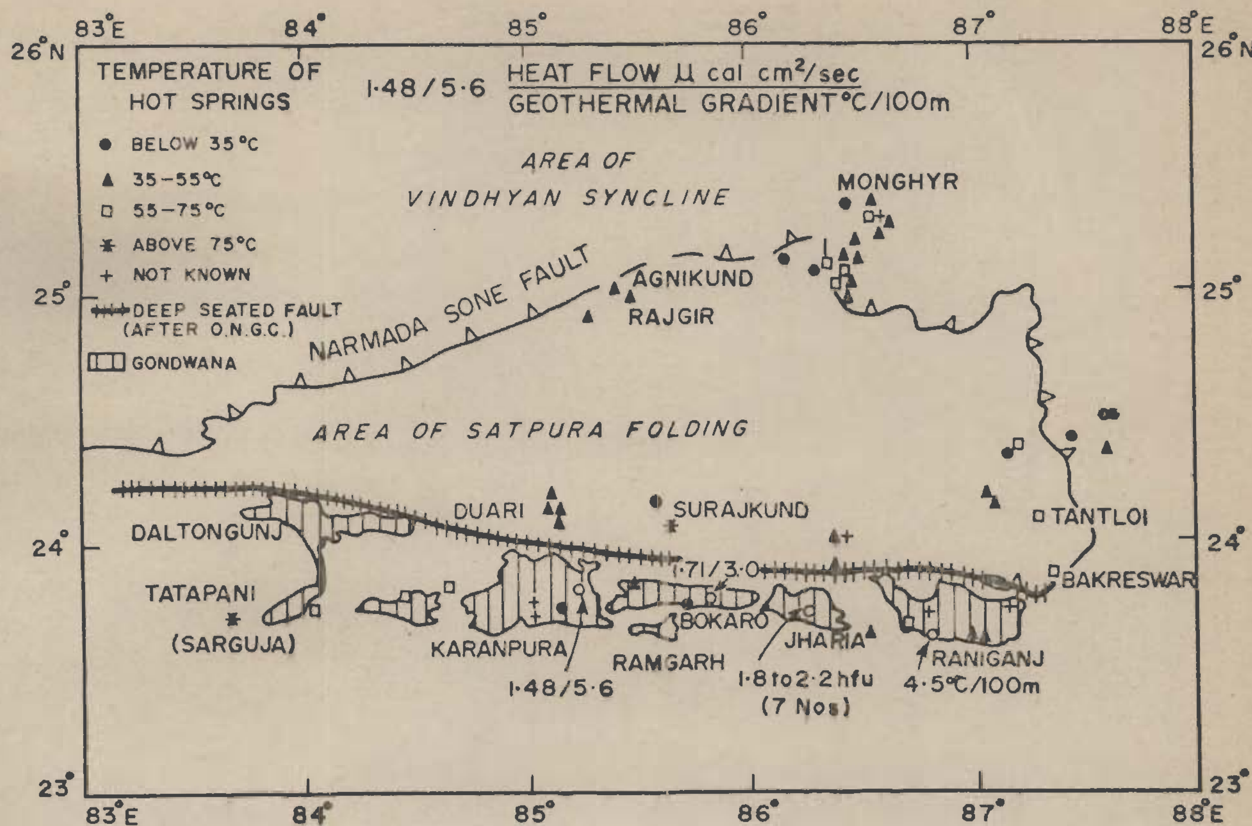


FIG. V-4 Hot springs of north-eastern part of India of Delhi apart from a large number widely distributed in other Pre-Cambrian fold belts of the Indian shield in areas of minor-faulting and shear zones, (Figure V-1).

5.2.1 Geochemistry of thermal waters

Geochemical data on hot spring waters supply useful informations regarding the geothermal resources potential (section 2.5.5). The geochemistry of the thermal waters of India has been considered by Chaterjee in 1969 and Gupta et al.(1975a). Detailed data of chemistry of thermal waters of various regions are also available, Deb & Mukherji (1964); Gupta and Saxena (1974, 75, 79); Handa (1975), Jangi (1976). All the known data has been re-examined so as to get a better insight into the geothermal resources potential of India.

As regards Himalayan Orogen partial and in some cases detailed chemical information of its thermal waters of various hot spring areas is available, and thermal waters of nearly neutral to mildly alkaline Na-HCO_3-Cl, Na-Ca-Cl-HCO_3, Na-Cl- Na-Cl-HCO_3, Na-HCO_3, Ca-Mg-HCO_3, Na-Mg-HCO_3(mixed type) types are known to occur. An analysis of data and comparision of weight ratio of thermal waters in volcanic hot springs as reported by White and others in 1963 indicate association of some magmatic components in some thermal waters of the Himalaya. To site an example, Puga thermal waters (section 6.5) are dominated by relatively high F, Cl, SiO_2, B, CO_2 and Li, as well as, low Ca and Mg and also have large (11-12 mg/lit) Cesium. This indicates association of some phase/s of late magmatic activity in the area. Association of these waters with borax and sulphur also indicate contribution from magmatic bodies as sources of heat and some of the solutes.

Most of the hot springs in the Himalayas occur in river valleys. It is likely that those manifesting in a particular river valley share a single hydrothermal system. Chemical similarities if observed in thermal waters of a river valley point out towards such a situation (section 6.8.1 and 6.8.7). It has been found that generally the hydrothermal systems of Himalayas are characterised with base temperature of the order of 150°C or over (Gupta et al. 1975a, 76c, 79), Table I-3). Thus indicating good geothermal resources potential in the Himalayas.

Next important hot spring region is the Konkan Coast, which is a strip of land of about 50 km wide lying between the Western Ghats and the Western Coast of India. Its thermal waters are dominated by Na^+- Cl^- and Ca^{++}- SO_4^{--} ions. Most of the springs in this region issue from the same geotectonic environment and indicate more or less similar chemical character of thermal waters, thereby suggesting a cognate source. However the spring waters lack in appropriate concentrations of boric acids, fluoride, ammonia, hydrogen sulphide, which are found in significant quantities in waters of high temperature geothermal areas. The main gasses in such areas are CO_2 and H_2S. Analysis of gases discharged from some of the main hot springs of Konkan has shown low contents of CO_2, H_2S and

H_2. Thus suggesting that contribution from high temperature rocks (magmatic sources) to the gas discharge are negligable. It appears that low enthalpy fluids can be tapped in the Konkan Coast.

Next regions worth consideration are the various grabens of the Indian land mass; viz., Cambay, Narmada-Sone, Godavari, Damodar etc. Cambay Graben is well known for its hydrocarbon resources. Highly mineralised (about 4.6 g/lit) thermal waters were found to flow from two exploratory oil wells in the graben. Fluid chemistry of thermal waters and those stored within the sediments at similar depths from which thermal waters had been tapped, indicate a different source of the high enthalpy fluids, which probably lies within the traps. Hot springs are only found near the periphary of the graben. However in the case of Damodar, Mahanadi, Godavari etc., Grabens, hot springs are known to emerge through the Gondwana sediments in various locations, but detailed chemical analysis of their waters, except for few locations of Godavari valley, are not available. Reservoir temperature around 160°C has been estimated for one hydrothermal system (Agnigundala) of Godavari valley (Saxena and Gupta, under preparation).

Hot springs are wide spread and manifest through a variety of rock formations in Precambrian fold belts. Their thermal waters differ significantly in physico-chemical characteristics. The total mineralisation is low and varies from

0.

0.06 to 0.7 g/lit. A detailed search indicate that the known mildly acidic waters of India issue through the Archaean quartzite in Rajgir-Monghyr belt of hot springs where movements have occurred during Miocene to Pleistocene. The hot springs of this belt although distributed over a large area, separations between two hot spring areas can be as large as 150 km, show the same chemical character and have very low mineralisation 0.06 g/lit. The waters differ from the local ground water in their chemical character and their origin is difficult to explain. They are in fact rather unique, being radioactive and associated with 15 to 20% CO_2; 1% CH_4; 6 to 9% O_2 and 70 to 80% N_2 and inert gases (Guha and Nag, 1971). CO_2 is generally a major constituent of gaseous material from areas of volcanic hot springs, and contribute from 70 to 95% of the total volume (White, 1957; Mahon, 1970). The low and high contents of CO_2 and N_2 respectively in the spring gases of the belt indicate that their volcanic/magmatic origin is very remote possibility. Thus suggesting to their very low geothermal potential.

Another group of hot springs with temperatures varying from 35 to 91°C (Figure V-4) occurs towards the south of the Rajgir-Monghyr belt through the Pre-Cambrian Chotanagpur genissic complex. The thermal waters are alkaline, of mixed type with TDS 0.6 g/lit and mostly dominated by $Na^+HCO_3^{--}$. Considerations of chemical parameters and other environmental factors point out their meteoric origin. However the hot springs of Tatapani (Figure V-4) (Surguja district) which has a discharge temperature near 91°C and some sinter deposits in its vicinity needs further attention.

Most of the hot springs of the Pre-Cambrian belt emerge through fractures at points which are lowest topographically. Rocks of various types with little or no primary porosity or permeability occur in their vicinity. The effective porosity and permeability being due to fracturing, shearing or jointing. In the peninsular shield, various old faults have been delineated and several earthquakes of small magnitude recorded during the last decade. The northward movement of the Indian plate, which is still continuing and its collision with the Eurasian plate is a major tectonic activity of far reaching consequences. Associated phenomena with this and minor earthauake activity, have kept the old faults and fractures open and thus facilitated the deep circulation and heating of meteoric water and emergence of hot springs.

On the whole, chemical parameters alongwith other considerations indicate that the waters of the hot springs, which occur in various Pre-Cambrian fold belts of the Indian shield, are mostly of meteoric origin. Their temperatures are controlled by the geothermal (mostly normal or below normal) and hydrogeological regimes of the respectiver regions.

5.4 Thermal Field of Indian Landmass - A synoptic View

Observed geothermal gradients in rocks of various ages and types in India (Figure V-5) show a large scatter and variation. The highest temperature gradient (45 to 75°C/km)

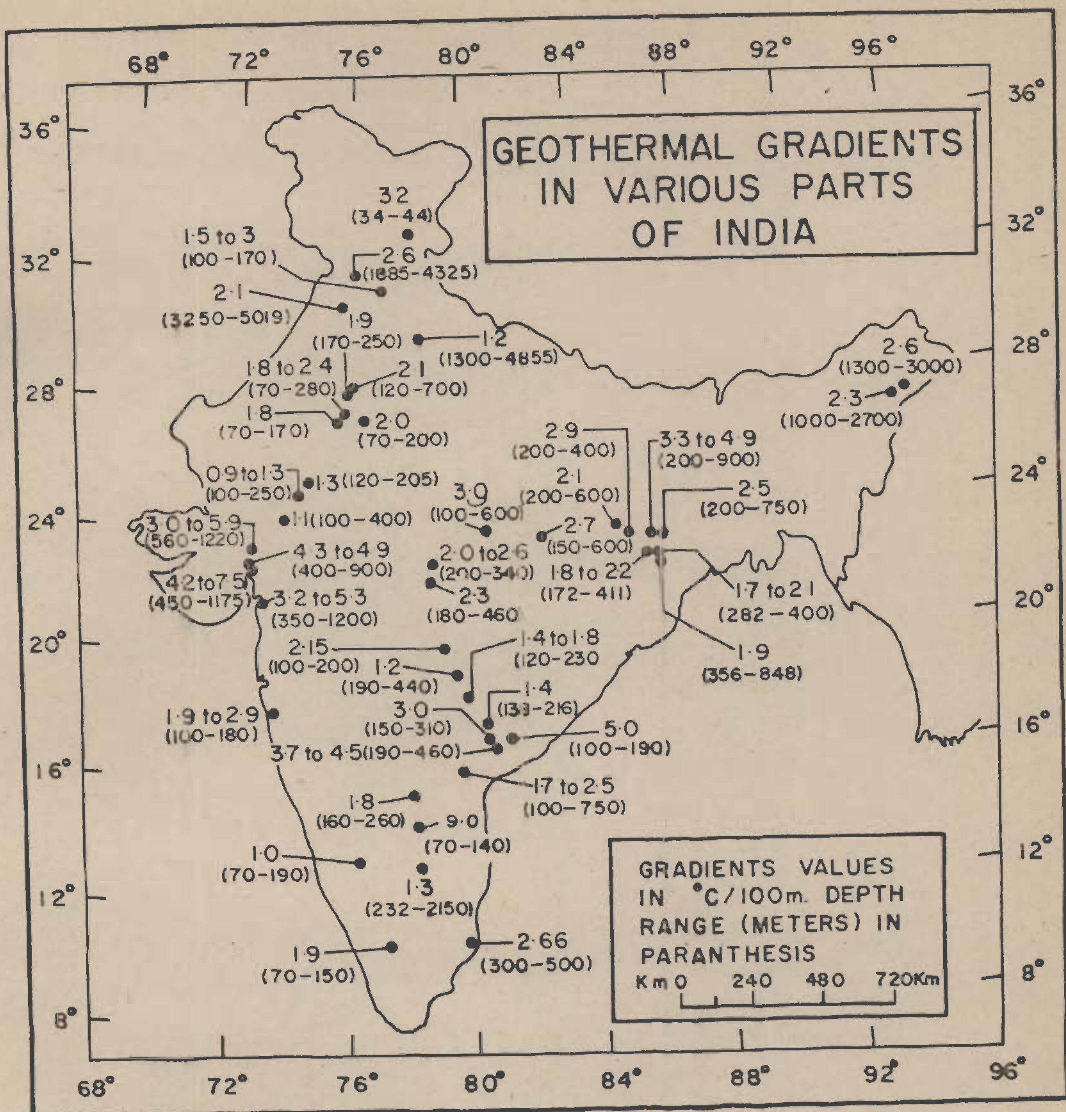


FIG. V-5 A sampling of geothermal gradients in India so far observed has been found in the Upper Tertiary hydrocarbon bearing sediments which overlie the Upper Cretaceous - Eocene flood basalts. Next in line come the coal bearing Gondwana sediments showing somewhat lower gradients (27 to 60°C/km).

The observed heat flow in the Cambay graben range from 1.3 to 2.3 HFU. A zone of high heat flow in the middle part of the basin, located towards north of Mahi river, has been also delineated. The cause of high heat flow has been attributed to an igneous intrusion in the crust beneath the basin during the Pliocene-Miocene times vide section 3.6.

The heat flow values so far measured in the Damodar Gondwana Graben which is believed to be a rift valley, show a large variation, ranging upto 2.45 HFU. However, a systematic analysis of the heat flow data after duly accounting for the disturbing factors which cause large variations in the heat flow field, give a value of 1.8 HFU. This value of heat flow is somewhat higher than the global average. Recent uplift with noticeable arching of the crust indicating recent mantle disturbances which has been reported in the Damodar Graben, may be the likely cause of high heat flow.

Pre-Cambrian Terrains of India, as elsewhere are characterised with low geothermal gradients (9 to 24°C/km). However, some portions of the Indian Pre-Cambrian shield are associated with significantly higher surface heat flow as compared with the world average.

The thermal field of the Himalaya which, geologically speaking, attained its present landscape configuration only about a million years ago and where rock suites dating from recent times to the Pre-Cambrian are exposed in all of its four contrasted lithotectonic units, is certainly of a very complex nature. The Himalayan region has experienced compression buckling, heating and upward movements eversince the collision of the Indian plate with the Eurasion plate. Arched block warping of such highly heated crustal zones lead to lowering of pressures, which can locally generate magma. Emplacement of post-Orogenic granites in the Himalayan region probably occurred in this way. Small intrusive masses are indeed expected to have been emplaced in the Himalaya in the wake of such processes (Figure V-6).

Measurements at 3 locations in the Himalayan foot hills revealed normal temperature gradients of 26-32°C/ km and heat flow values of 1.2 - 1.3 HFU, but the Great Himalaya located between the Main Central Thrust and the Counter Thrust are characterised by high heat flow with some localised pockets for very high heat flow.

5.5 Geothermal Provinces in India

On the basis of the aforesaid analyses of available data, the four main geothermal provinces of India which have been selected for detailed studies are:

i) The Himalayan-Burmese-Andaman Nicobar Arc Geothermal Province.

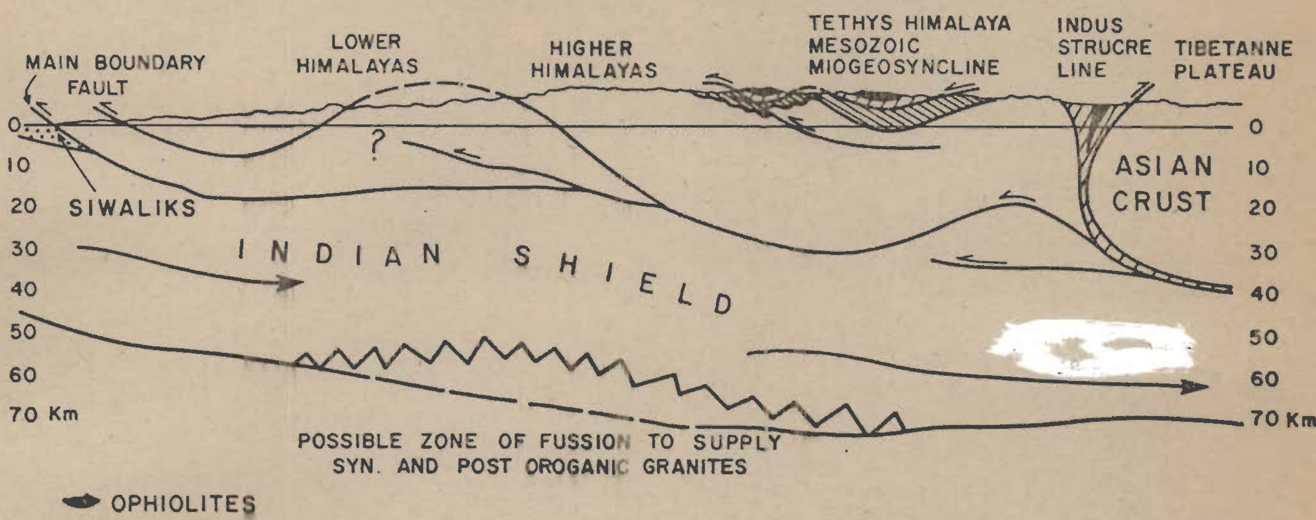


FIG. V-6 Schematic cross-section, south to north, through the Indian Shield, depicting collision with the Asian plate

ii) The Narmada-Sone-Dauki lineament Geothermal Province.

iii) The Konkan Geothermal Province.

iv) The Cambay Graben Geothermal Province.

Additionally, there are a few smaller geothermal provinces where hot springs and thermal waters of low enthalpy are encountered. However, these are generally areas of normal geothermal gradient where the meteoric waters have been able to penetrate to great depths and return to the surface thereby transporting the geothermal energy. At some places some exothermic chemical reactions might also play an active role. However geothermal resources of such provinces (with two exceptions: Agnigundala and Tatapani-Surguja) are not promising for exploitation from the techno-economic point of view.

5.5.1 Himalayan-Burmase-Andaman-Nicobar arc geothermal province

This province forms a part of the Alpide-Himalayan zone, which is a region mostly of continent-continent collision. Crustal shortenings of the order of 200 to 500 km appear to have occurred in this zone during the last 10 m.y. The zone of intercontinental shortening runs from Iran through Baluchistan, Himalayan and the Burmese Arcs to the Andaman-Nicobar Islands. A large part of the differential motion is probably accomodated along the Zagros fold belt, the Baluchistan and the Himalaya arcs (Lepichon et al. (1973).

The Himalaya were largely formed by middle to late Miocene times (10 m.y.). However, geologically speaking, the Himalayan mountains attained their present configuration only about a million years ago. This mountain belt, bordering the Tibetan plateau to the south, extends as an arcuate belt for over 2,500 km, flanked by remarkable syntaxial bends on its western and eastern ends. The eastern bend joins it to the Burmese ranges (Arakan Yoma) which continue upto the Andaman-Nicobar Islands through a submarine ridge.

Earthquake activity in the province is not uniform all along the arc. Clusters of epicentres occur in the Hindukush, Northwestern and Western Himalay and in the north-eastern part of the country down to the Andaman-Nicobar Islands (Figures V-2 and V-7). Most of these earthquakes are of shallow origin except in the Hindukush region where they occur at intermediate depths. Few intermediate depth earthquakes also occur near the Indo-Burma border region and the Andaman Sea.

A belt of hot springs extends from NW Himalayas through Nepal, Bhutan to NE Himalayas towards Burma and finally to the Barren Islands (Fgiures V-2 and V-7). Most of these are located in the north-western part of the Himalayas while a few are known to occur in Sikkim state and Nepal and Bhutan. More detailed studies may, however, reveal additional hot springs in the eastern part (Gupta et al. 1975). The Geological Survey of India has reportedly located some centres

of hot

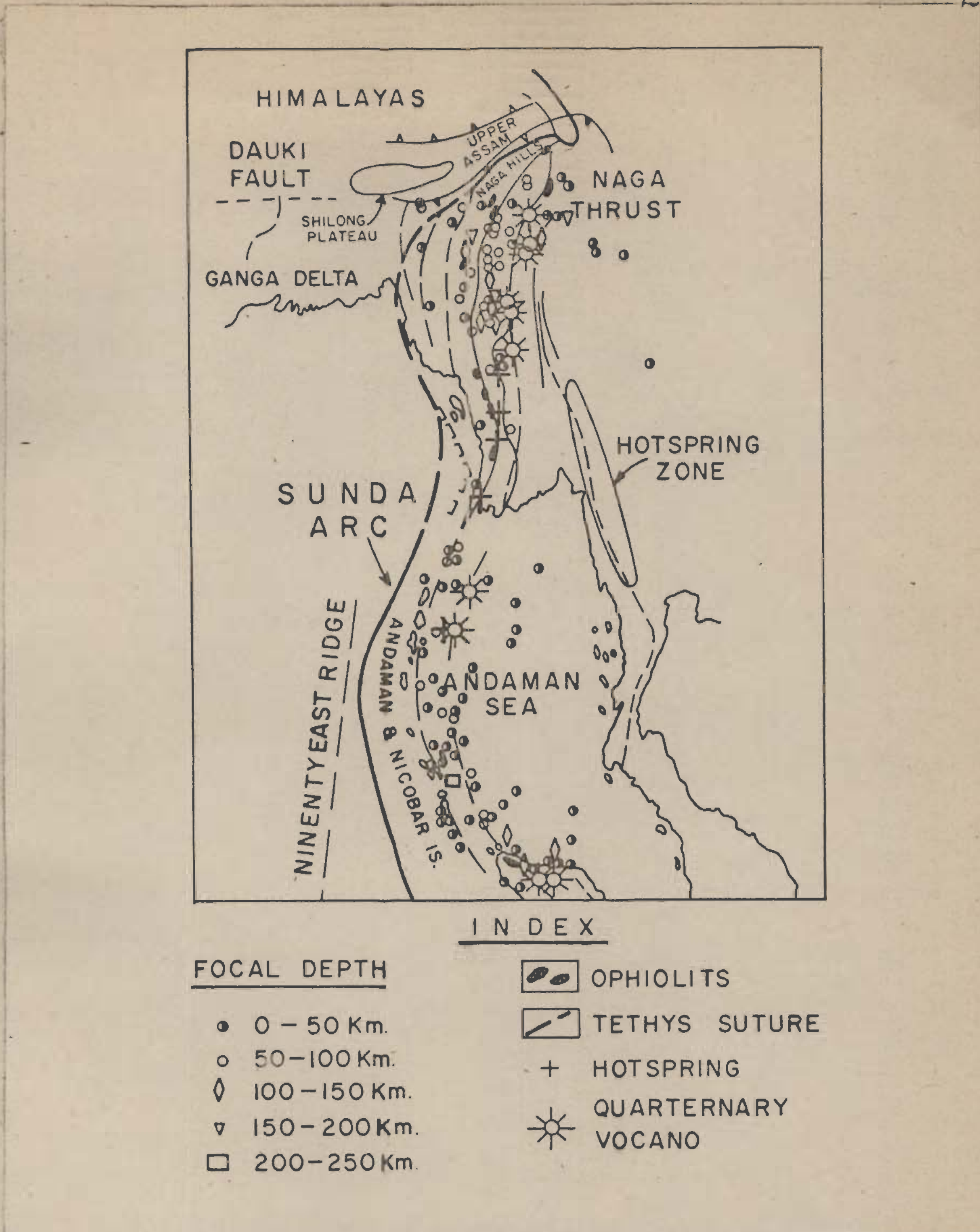


FIG. V-7 Tectonic features near the eastern India hot spring zone of hot springs in the Assam-Nepal region. In the Barren Island which is famous for mud volcanoes only one group of hot springs is known to occur.

5.5.1.1 Sub-provinces

This rather large and sprawling province can be divided into two main sub-provinces: i) The Himalayan Geothermal sub-province and ii) the Burmese-Andaman-Nicobar arc Geothermal sub-province. The former can be further sub-divided into smaller sub-provinces, namely the NW Himalaya and the Eastern Himalaya geothermal sub-provinces. However till more detailed studies are available and discriminating characters and geoparameters have been clearly observed, the Himalaya may be considered as a single geothermal sub-province.

5.5.1.2 Himalayan geothermal sub-province

Attempts to construct the evolutionary and tectonic history of the Himalaya have been made since the 1930s by various workers, beginning with Auden in 1934 and 1937, Heim and Gansser in 1939, West in 1939, Wadia in 1962, Pande and Saxena in 1968, and Valdiya in 1969 (See Valdiya, 1973). These studies have shown that the Himalayan tectonic belt consists of four contrasted lithotectonic units, as shown in Figure V-2. These units from south to north are:

1. The outermost Himalayan unit consisting mostly of Siwalik sediments of Miocene-Pleistocene age, which form hills of varying widths.

2. The Lesser Himalayan sedimentary zone consisting of folded and faulted Precambrian to younger sediments.

3. The Great Himalayan Central Crystalline Zone, containing practically all the snow-clad high peaks, consisting of some sedimentary and metamorphic rocks and large masses of igneous intrusions.

4. The Thehya-Himalayan Zone which is composed of sediments of all ages formed in the Tethyan geosyncline.

All four units extend from east to west in continuity and are sharply delineated by tremendous boundary thrusts. Figure V-2 clearly shows that in many areas, these Himalayan thrusts are sesimically active.

The northern boundary of this sub-province is separated from the Karamoram-Kailas ranges by a great counter thrust called the Indus Suture Line (belt by Gansser, 1964, 1966). This suture belt which resulted from the fracturing of the northern boarder of the Indian Peninsula when it collided with the Eurasian plate (Tibetan Block), is a narrow belt of deep fracture (50 - 60 km) located along the Indus and Brahmaputra rivers. It is composed of strongly compressed (sandwitched between the Himalayan belt of rocks to the south and the Ladakh granite towards the north in N-W Himalaya) Upper Cretaceous to Eocene continental deposits of Indus flysch and associated ophiolites. According to Tiwari (1964) the Indus flysch represents a molasse facies and are probably of post-Eocene to Miocene age. The main large scale thrusting which caused the squeezing of the Indus flysch and associated ophiolites also caused N-S cross faulting both in the Indus zone and the Ladakh zone. The cross faults developed as relief structures following large scale thrusting, Shah et al. (1976).

Another thrust of deep fundamental character of this subprovince is the Main Central Thrust, which continues to be active, Valdiya (1973). This thrust along with its associated faults and thrusts, separates the lesser Himalayan sedimentary zone in the south from the metamorphics of the Central Crystalline Zone in the North.

As explained earlier magmatic hearths can locally appear in highly heated zones of the crust experiencing arched block-warping. These processes which occur quite commonly in the Himalayan Orogen, have created potential geothermal fields through out this province and even beyond the Indus suture line in the Karakoram-Kailas region. Youthful uplift in the Himalaya not only produces large fracture systems, but keeps these open, thereby facilitating the emergence of hot springs which usually manifest themselves along the faulted contacts of Tertiary and Pre-Tertiary sediments with granites and gneisses, and in some cases through joints and faults in the crystalline rocks. Misra and Bhattacharya (1976) observe that in the Central Crystalline Zone of northern Kumaun Himalaya, a number of springs and prominent nalas have been found to originate, all around the thrust plane.

Whilst only a few warm springs are found in the Lesser Himalayan Sedimentary Zone and the Siwalik sediments, they occur more frequently further north e.g., the high temperature hot spring of Puga, Manikaran, Jumnotri, Tapoban etc., all occurring in the Great Himalaya, usually close to one of the bounding thrusts. Boiling with steam separation and mild geyser activity in India is only observed in this region. Various hot springs including the high temperature spring of Nubra (Figure V-1) also emerge in the Karamoram-Kailas region towards the north of the Indus-Suture Counter Thrust.

Hot springs located in between and in the vicinity of the Counter Thrust and the Main Central Thrusts are usually found to possess good geothermal potential as recent investigations reported in a later chapter indicate.

5.5.1.3 Probable heat sources of hot springs

The Himalayan belt is not marked by any recent volcanic activity, which is true also for the rest of the Alpide-Himalayan tectonic belt from Assam to Central Europe, along which many hot springs occur. In the Himalaya there has been and continues to be intense tectonic and orogenic activity, along its numerous faults and thrusts. Heat is normally generated by orogenic activities, by the disintegration of radioactive elements, and by exothermic chemical changes. Heat is given out by cooling of uplifted rock masses. Preferential heat transport occur through tectonic fabrics such as foliation or schistosity perpendicular to the stratification, etc. But neither of these can, alone or collectively constitute geothermal resources which would be economically viable.

Whilst thermal waters obtained from such environments can be profitably exploited for various uses such as space heating, agriculture, etc., basically, it is the heat conducted away from shallow hot magmatic bodies along which can provide a sustained economically viable source of geothermal energy.

Granitic intrusion in the Himalayas has already been discussed in section 5.2. Granitic and granodioritic rocks ranging in age from Precambrian to Uppermost Tertiary, are predominently confined to the Great Himalayan Central Crystalline Zone. Large granite emplacements were produced during compression and uplift of the region through processes discussed earlier. The Tertiary granitic rocks predominate over the rocks of other ages. For example the Badrinath-Kedarnath granites are of post-Miocene age (Gansser, 1965). Valdiya (1973) suggests post-Tectonic emplacement of Badrinath granites (possible 18 m.y.). Recent investigations in the Puga-Chumatang region show evidence for the occurrence of Miocene/Pliocence granites in the Ladakh range near Chumatang. The granite intrudes into younger Indus formations which are post-Eocene-Miocene, not as old as Cretaceous (Tiwari, 1964). Thermal waters of Puga show evidence for their association with some late phase/s of magmatic activity. Granites of Tertiary age exposed in the vicinity of some Himalayan hot spring areas could be possible heat sources if they are really younger (post-Miocene).

There are also granitoids in the lesser Himalayas. These bear a resemblance to granitic rocks of the peninsular type. These are in all probability the result of involvement of the northern margin of the Gondwanaland in the Himalayan orogeny. Such granitic masses, though very rare, are the result of northward under thrusting of the northern margin of the Gondwanaland.

It is therefore very likely that granites formed at various times due to one or other factors outlined above are the source of heat for most of the hot springs of the Himalayan orogen.

5.5.2 Narmada-Sone-Dauki lineament geothermal province

The Narmada-Sone lineament and the Dauki fault in the eastern part of India mark a most prominant east-west feature line of the sub-continent. West (1962), Ahmad (1965), and Choubey (1970) have emphasized the fundamental nature of the Narmada-Sone line. Extended eastward, this line joins up with an active normal fault -- the Dauki fault -- which borders the Khasi-Jaintia Hills in the Shillong plateau. The Dauki fault then merges with the great Naga thrust of eastern India. To the north of the Narmada-Sone-Dauki line lies the Vindhyans and their equivalents in the Lesser Himalayas; to its south, stretch the Gondwanas in linear basins formed in the Cuddapah and Archaean floors (Jhingran, 1970). It is significant that south of this line the Pre-Cambrian Vindhayns do not occur before the Wardha valley. North of this line the Deccan trap lavas rest directly on the Vindhyans; to the south they rest on the Gondwanas or on the Archaeans, the Vindhyans being absent. West (1962) and Ahmed (1965) consider the Narmada-Sone line as a major fault zone of the Indian peninsula and probably an ancient feature reaching the deep mantle.

The succession of events in the Indian peninsula suggests that the line of the Narmada-Sone valleys has been a line of crustal weakness since the Precambrian, and has been repeatedly activated by crustal movements operating intermittently along this line. On the basis of combined structural and geomorphological observations, Choubey (1970) concluded that rift-faulting has occurred along this line. Bhimasankaram and Pal (unpub. data) on the basis of palaeomagnetic traverse taken across the Narmada, speculate it to be a strike-slip fault. Choubey (1970) suggested that this ancient feature has been connected in someway with primary weak zones parallel to the Archaean grain. Naqvi, Rao and Narain (1974) postulate the merger of the Singhbhum and Dharwar protocontinents with the Aravali protocontinent along the Narmada-Sone line. If this line of merger is extended towards the east, it would seem to fall in line with the Dauki fault. The Narmada rift is also bordered on the north by a line of carbonatite intrusions (Yellur, 1968, as quoted by Jhingran, 1970). Jhingran postulates that in the pre-drift configuration of Gondwanaland the Narmada rift lay in continuity with the African belt of carbonatite along the Mid-Zambesi-Laungawa rift of central and East Africa.

The east-west trending Dauki fault demarcates the Shillong plateau against the sedimentary plains of Bangladesh. This fault merges with the Naga-Dismog thrust on the east. It is most probably normal fault along which the Assam Plateau has been uplifted as a horst. Murty (1970) postulates an upward rise of the mantle material in the eastern part. This would result in the rise of isogeotherms; where this rise is steep, as in the Himalaya the deformation would be accompanied by regional metamorphism, with formation of synkinematic granites. A system of east-west trending vertical faults has also been mapped in the areas; these are parallel to the Dauki fault (Sengupta and Khatri, 1973).

Various hotsprings are located in the Narmada-Sone valleys, and towards the south in the Damodar graben and also at the margin of the Shillong plateau (Figure V-1). The geochemical character of the hot springs of this province, except for some located near the Damodar Graben, is not known. The Narmada-Sone-Dauki line zone extends beyond these limits westward to the Saurashtra peninsula and eastward to the Naga thrust. This ancient rift, possibly analogous to the East African Rift, is a deep fault which divides the Indian subcontinent into atleast two major blocks (Jhingran, 1970). Movements have continued in this zone and it is a pretty likely place for the concentration of heat energy to form potential geothermal energy sources in the region.

5.5.3 Konkan geothermal province

The Western Ghats -- a chain of mountains running parallel to the western coast of India in a north-south direction, at an average distance of about 50 km from the shore -- border a coastal strip of land known as the Konkan area. A large number of hot springs or group of hot springs with maximum temperatures upto 70°C emerge from the Deccan trap in this belt between latitudes 16°N and 20°N (Figure V-3). The Deccan traps consist of a number of flows, covering about 512,000 sq.km. It is generally accepted that a fissure type volcanic activity continued for a fairly long time, probably from the Upper Cretaceous to the Eocene. However, there are indications that volcanic activity on a small scale might have continued upto the Pleistocene, long after the close of the Deccan trap activity (Gupta et al., 1968). All the tectonic trends in the Konkan coastal tract are oriented in the N-S and NNW-SSE direction. The hot springs show a linear north to south spread. This fact along with the occurrence of lower to Mid-Miocene strata along the west coast is indicative of a set of faults parallel to the West Coast fault, which was suggested by Oldham (1983, p.493). Krishnan (1968) states that the remarkably straight edge of the western coast is indicative of a fault formed in the Miocene, running from south of Cape Comorin to near Karachi (Pakistan). The extent of down faulting being of the order of 2,000 m. Krishnabrahmam and Negi (1973), on the basis of the conspicuous axes of gravity lows, suggest the existence of two rift valleys beneath the traps;

the Koyna rift, passing through Koyna and covering a length of about 540 km, and the Kuruduvadi rift about 390 km long apparently merging with the Koyna rift near Poona. The average width of the rifts on the basis of gravity profiles appears to be about 50 km. These rifts are still active as indicated by earthquake activity.

A striking feature is that most of the hot springs of the Konkan are located near the axis of the Koyna rift. Another interesting feature is the location of many of the hot springs on the borders of dolerite dikes, which run in a north south direction near the hot springs. At most places the hills at the foot of which the hot springs emerge are found to have a triple flow formation, which is indicative of deep fissures in the area. However, general association of hot springs with topographic lows, as indicated by the presence of rivers or small rivulets adjoining the spring, is observed. Occurrence of frequent shallow seismic activity is also well known in the Konkan belt. The above mentioned factors are indicative that strong crustal movements occurred in the region and the fracture system which control the hot springs are still active.

5.5.4 Cambay graben geothermal province

The Cambay graben (Figure V-1) is located in the alluvial plains of Gujarat State of Western India, extending approximately from latitude 21°N to 24°N and longitude 71°31'E to 73°4'E. It is a Cenozoic basin and runs as a narrow graben in an approximately north-northwest to south-southeast direction. South of latitude 21°45' it takes a swing towards north-northeast to south-southwest and runs into the Gulf of Cambay. Extensive geophysical work in connection with oil exploration has been carried out in the basin since 1958. Geology, tectonic and heat flow of the basin has been discussed in detail in Chapter III.

It has been shown that high temperature gradients (upto 75°C/km) and high heat flow values varying from 1.80 to 2.3 HFU have been observed in the area located north of Mahisagar river. Two wells drilled for oil exploration in this part of the basin tapped steam-water mixture under very high pressure (50 to 100 atmospheres) and at temperatures over 100°C from depths of 1757 m and 1958 m. Discharge of 2500 to 3000 m^3/day have been estimated. The most probable source of the large volume of high-pressure thermal fluids must be some deep-seated reservoir in the Deccan traps where it must exist in the water phase. Great faults (possibly penetrated by the wells) and extending to the traps must have provided the channels for the upflow of thermal water. The above factors indicate that a potential deep geothermal reservoir lies in the Cambay graben. It is worth pointing out that high temperature (110°C) has been observed at a depth of about 50 m in a borehole drilled in the Towa hot spring area (Rao, Ghosh and Bhalla, 1979). The Towa springs are located in a fault zone in pegmatites in the Proterozoic Erinpura granite and are near the northeast margin of the Cambay graben. The existence of a connection between the heat sources for the thermal water of the Cambay graben and for the hot springs of the Towa region is doubtful. This only indicates that the areas near its margin, are, probably, also thermally anomalous.

5.6 Conclusion

It is clear that the analysis of various geodata, mainly the geological and the tectonic history of various units of the Indian landmass and of its thermal fields, and the occurrence of hot springs and their chemical character, indicates that a few geothermal provinces of India occur along tectonically weak zones. Keeping in view the current techno-economic considerations, potential hydrothermal resources in India can be explored and exploited for various uses including power generation.

CHAPTER VI

EXPLORATION OF GEOTHERMAL ENERGY RESOURCES

"In the long run, however, a major contribution may well come from the heat of the rocks in the earth's crust. One might envisage using the atom bomb to blast deep holes inside the earth's crust. Water could be made to enter these holes and come up in the form of steam to run turbines."

M.S. Thackar (1961)

6.1 Introduction

Currently the pattern of power production and its futuristic plans and projections areundergoing a conceptional revolution. This is the direct consequence of the technological break throughs motivated by spiraling demands for electric power. Governments in various countries are funding and seriously planning for the development and exploitation of new sources of energy. Geothermal resources belong to this unconventional group. Attempts to explore and develop the geoheat sources not only of natural hydrothermal systems but also of dry hot rocks have been initiated in a number of countries of the world and intensified in those where some exploitation is already being done.

In India the first steps in this direction commenced with the study of the thermal state of the Indian crust, initiated by a few individuals which was subsequently intensified with the formation of an Inter Agency "Hot Spring Committee" by the Government of India in 1966. As a sequel to this, a few earth scientists and technologists compiled the available information on the hot springs of the country, carried out reconnaissance work in four geothermal areas and recommended some others for future investigations.

As a member of a team of the said committee the author first carried out a preliminary thermal survey of the Puga hot spring area (Ladakh), during 1967. The survey indicated the possibility of encountering hot water at a temperature near boiling at a depth of about 50 m, and of retrieving borax from the thermal waters and using the thermal fluids for refining the borax deposite at the site (Gupta, 1967). This was later confirmed when exploratory boreholes drilled in the area spewed out sulpherous hot water and steam/water mixtures. Spurred by these encouraging manifestations of a potential geothermal source, detailed investigations were carried out to establish its character and extent. The geothermal data and the results of field investigations so obtained i.e.,estimates of the natural heat loss from the surface, probable reservoir temperatures, subsurface thermal conditions of the Puga valley form the subject of this chapter. Additionally results of some investigations in the Parbati valley (Manikaran-Kasol) hot spring area are also included.

6.2 Puga Valley Hot Spring Area, Ladakh

6.2.1 Location and geology

The area under investigation lies in the Puga valley which is situated towards south-east of Leh at a distance of about 220 km from it in the Ladakh district, in the NW Himalayas at an altitude of 4,400 m above m.s.l. The valley located towards east of Samdo and trending E-W over a stretch of about 15 km with a maximum width of one kilometre is situated between the Zanskar and Ladakh ranges. It lies to the south of the Indus suture subduction zone. This zone with ophiolite and associated rocks represents the remnants of an uplifted wedge of the oceanic crust and presently lies compressed between two continental plates (the Indian and The Eurasian).

The geology of the area has been studied by various workers, De Terra (1932), Raina et al. (1963), Tewari (1964), Shankar et al. (1974). The main rock sequence in the area is the Puga formation of probable Paleozoic age. It is comprised of paragneisses, quartz-mica schists and phyllites interlayered with bands of limestone. The Puga formation which is sparsely seen on the slopes of the mountain ranges bounding the valley, is at places obliquely intruded by basic rocks and amphibole-chlorite-schist in the form of dykes and sills. The intrusions are thin and are exposed on the southern slope of the valley (Raina et al. 1963, as quoted by Baweja 1967). The Puga formation is intruded by Polokongkala granite in the west. In the east is exposed the Samdo formation, which comprises

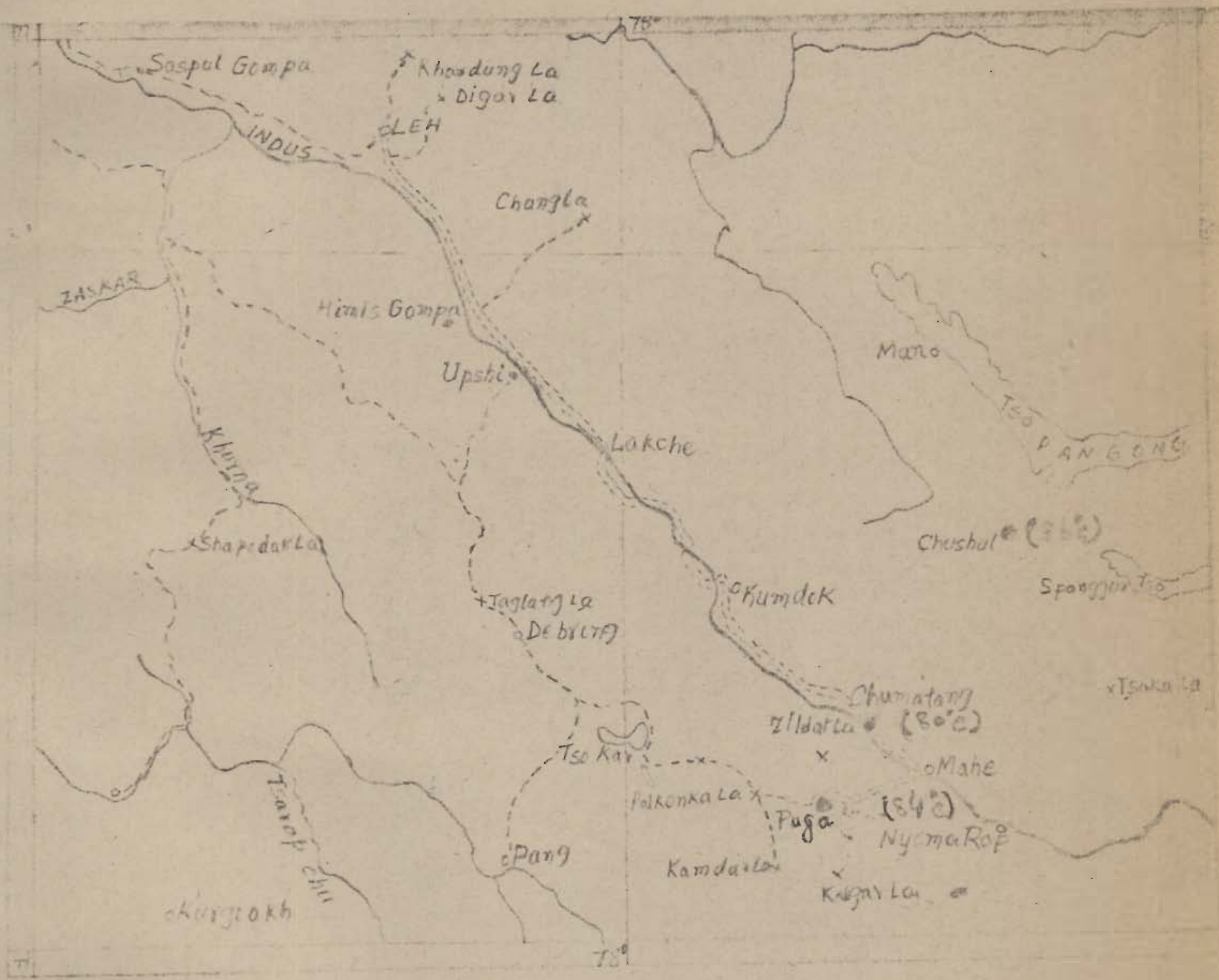


FIG. VI-1 Location Map of Puga Geothermal Field of volcanic flow, ash beds, tuffs and associated sedimentary rocks intruded by ophiolite suite (Shanker et al. 1974). Towards the north of Samdo formations and south of Ladakh range, a thick pile of clastic sediments called the Indus formation or Indus Flysch are present. According to Tewari (1964), this formation is a sequence of largely continental molasse facies of post-Eocene-Miocene age. This has been intruded by a granitic batholith - the youngest phase of plutonic and granitic activity (Post-Miocene to Pliocene?) in the area. The region therefore seems to have witnessed intense plutonic and sub-marine volcanic activity from Middle to Upper Cretaceous age and various phases of acid igneous activity from Upper Cretaceous to Upper Tertiary.

The Puga valley appears like a graben or a down-faulted block with its northern and southern faults concealed under the valley material. Detailed geological mapping (Raina et al. 1963, Shanker et al. 1976) has indicated that the Puga valley is aligned along the faulted crest of an asymetrical anticline in Puga formation. The regional trend of the foliation in these formations is NW-SE with a dip of 25° towards NE in the northern range and of 35° towards SW in the southern range. The northern limit of the Puga formations is marked by a major NW-SE trending fault, named as the Zildat fault. Certain NW-SE and N-S trending faults are also suspected to occur in the northern flank of the valley. A concealed fault along the base of the northern hill is inferred from the presence, in the central part of the valley, of the sulphur deposit representing an old line of fumarolic activity. The valley floor is covered with recent to sub-recent deposits of glacial moraines (partly lake sediments), aeolian sand and scree, borax, sulphur and other spring deposits. Borax and sulphur, which widely occur in the central part of the valley and are genetically connected with thermal fluids are deposited by capillary action. The surface is hardened, in the vicinity of hot springs, due to the cementation of scree and sand of the valley floor.

6.2.2 Surface manifestations

The surface manifestations of the geoheat are in the form of a large number of hot springs, over hundred, with temperatures varying from 35 to 84°C (boiling point of water at Puga altitude), and individual discharge upto about 5 lit/sec. Extensive patches of warm ground, mud pools, hot water seepage in Puga nala, sulphur condensates and borax deposits also occur in the area. Almost all the surface thermal activity is concentrated in a 4 km long stretch of the valley towards its eastern part. Apart from isolated hot springs, which emerge on either flanks of the valley, the hot springs usually occur in clusters along the Puga nala. Few hot springs are located in the nala bed itself.

Encrustations of borax are noticed around each spring. Sulphur rings formed after reduction of H_2S originating from deep thérmal waters occur at several places along with borax.

6.3 Thermal Investigations And Surveys

6.3.1 One-metre temperature surveys

Temperatures were measured at a depth of one metre by using calibrated thermistor probes especially designed for the purpose (section 2.4), over a grid at seperation of 50 or 96 m in the N-S as well as in E-W directions, covering a 6 km stretch of the valley.

Isothermal maps based on these measurements are shown in Figures VI-2, 3 and 4 from the western part of the valley towards its eastern end.

6.3.2 Shallow temperature gradient studies

Temperatures at 0, 20, 40, 60 and 100 cm depths were measured in specially drilled small diameter holes at a large number of locations, mostly in a N-S direction on various traverse lines. Temperatures at various depths for some locations are shown in Figures VI-5 and 6 and an isogradient map is shown in Figure VI-7. The shallow temperature gradients generally vary from about 1.0 to 10°C/m, with exceptions generally in the vicinity of hot springs (Figure VI-7).

Shallow temperature surveys were carried out more or less during the same months in two different years at two stages; i) during the period when exploration for sulphur deposits was going on and ii) during the period when geoinvestigations for the exploration of geothermal resources of the

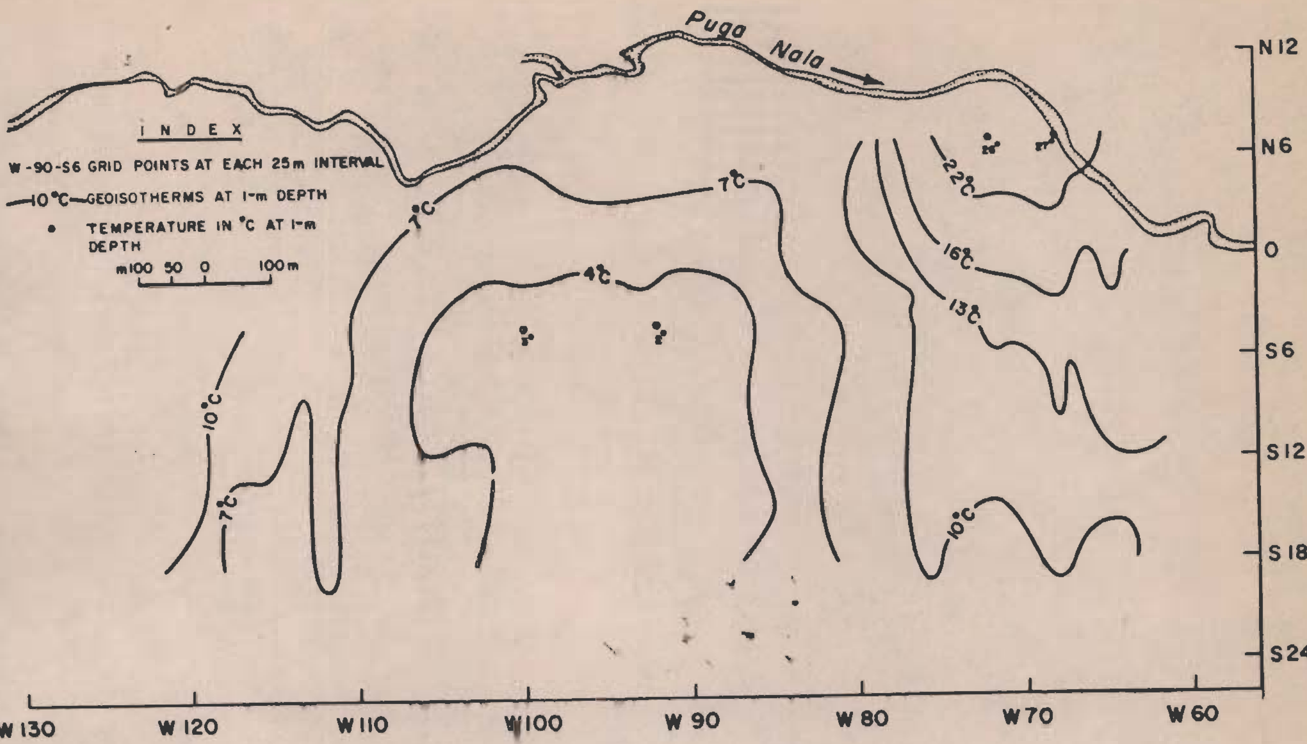
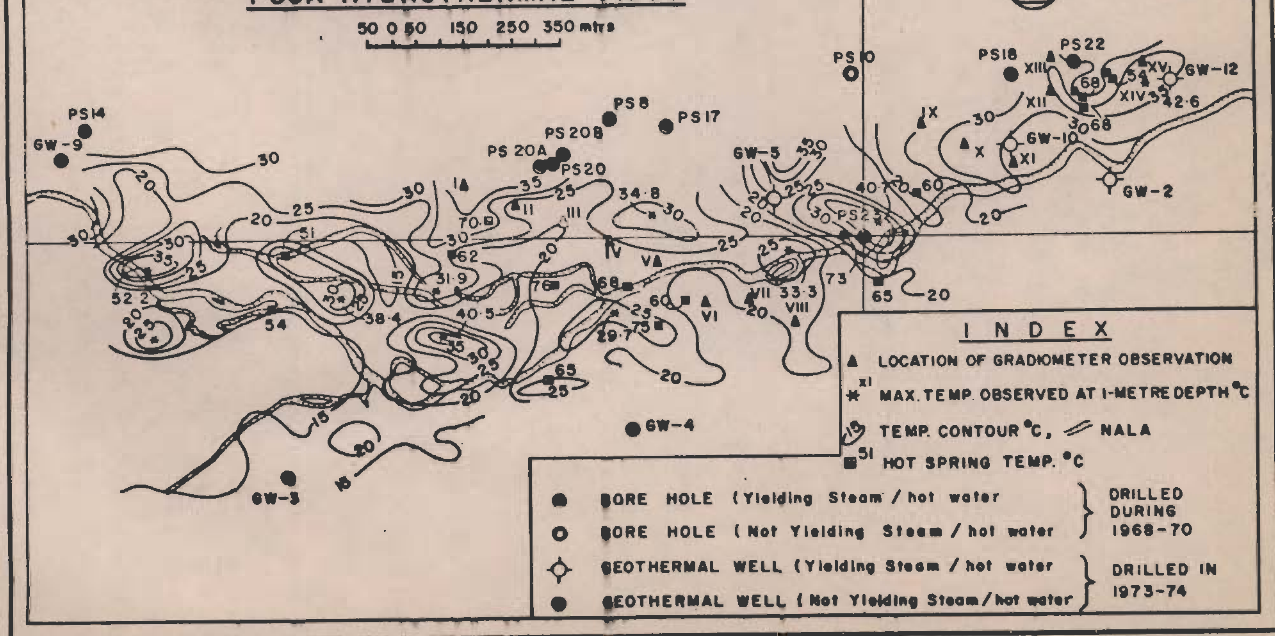


FIG. 2 GEOISOTHERMS AT 1-m DEPTH IN THE SOUTH-WESTERN PART OF THE PUGA VALLEY

Fig. 3- SHALLOW (1-METRE) TEMPERATURE SURVEY OF PUGA HYDROTHERMAL FIELD



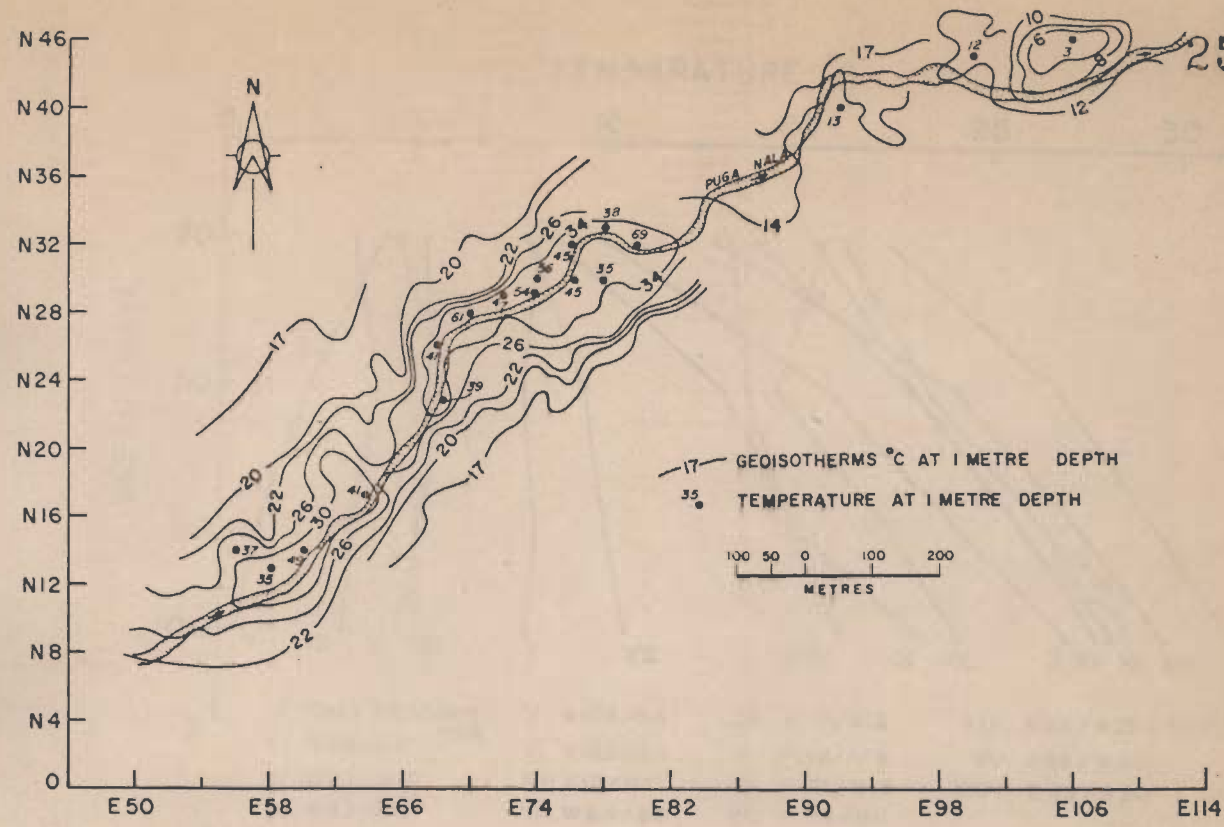


Fig.4- GEOISOTHERMS AT I METRE DEPTH IN THE EASTRN PART OF PUGA GEOTHERMAL FIELD

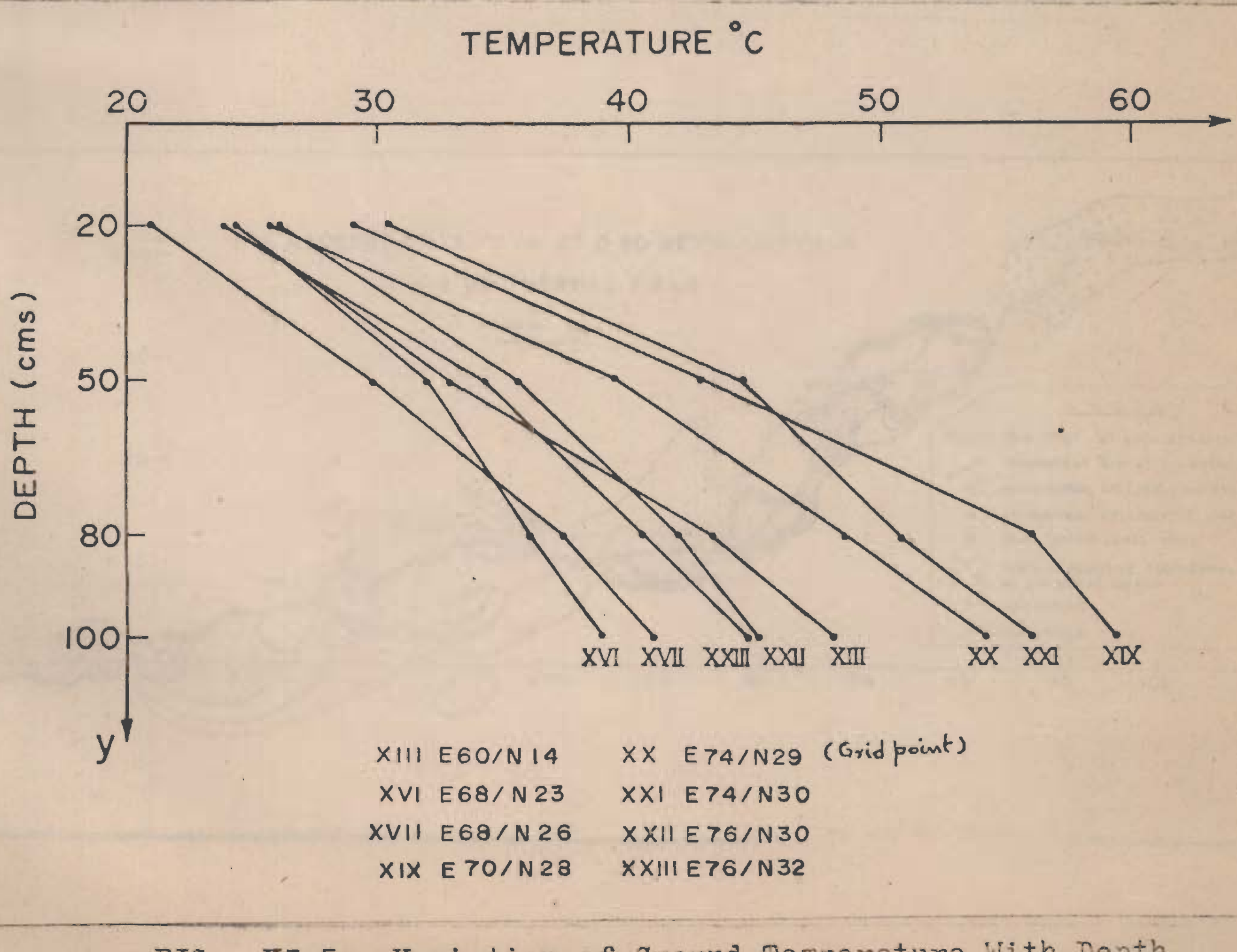


FIG. VI-5 Variation of Ground Temperature With Depth at Few Locations - Puga Geothermal Field

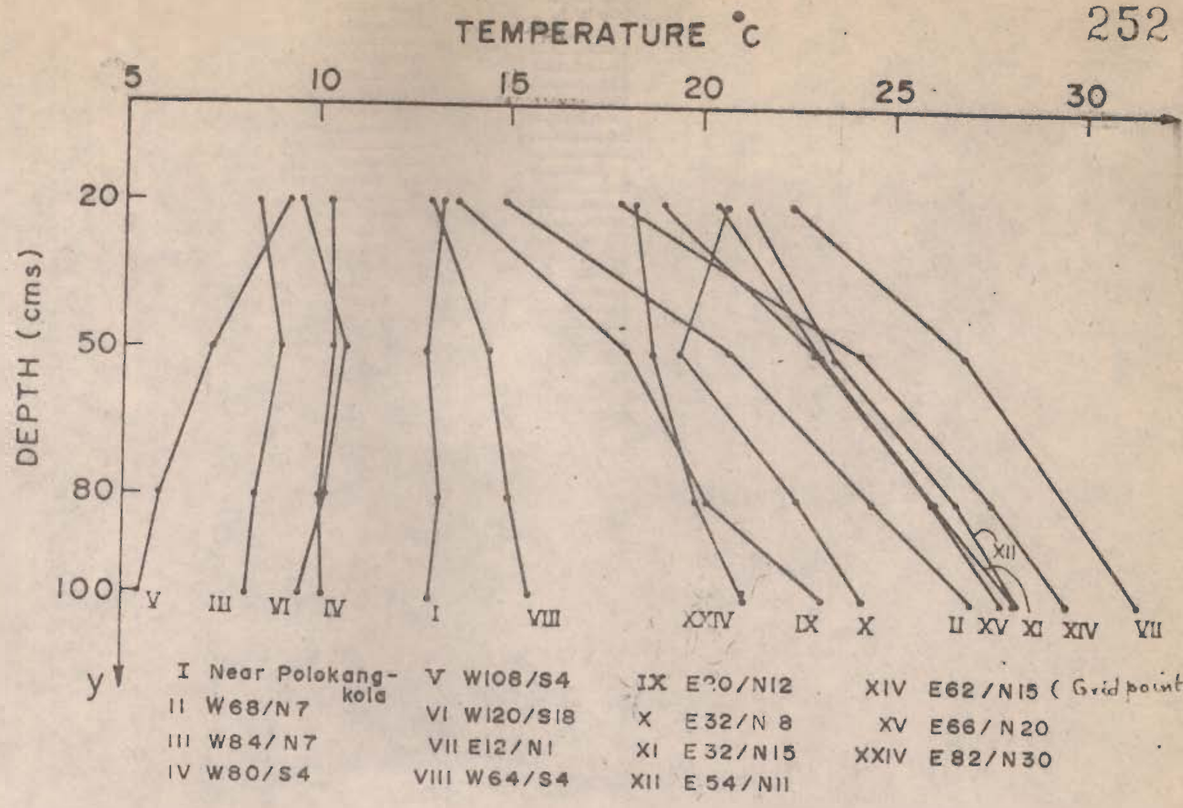
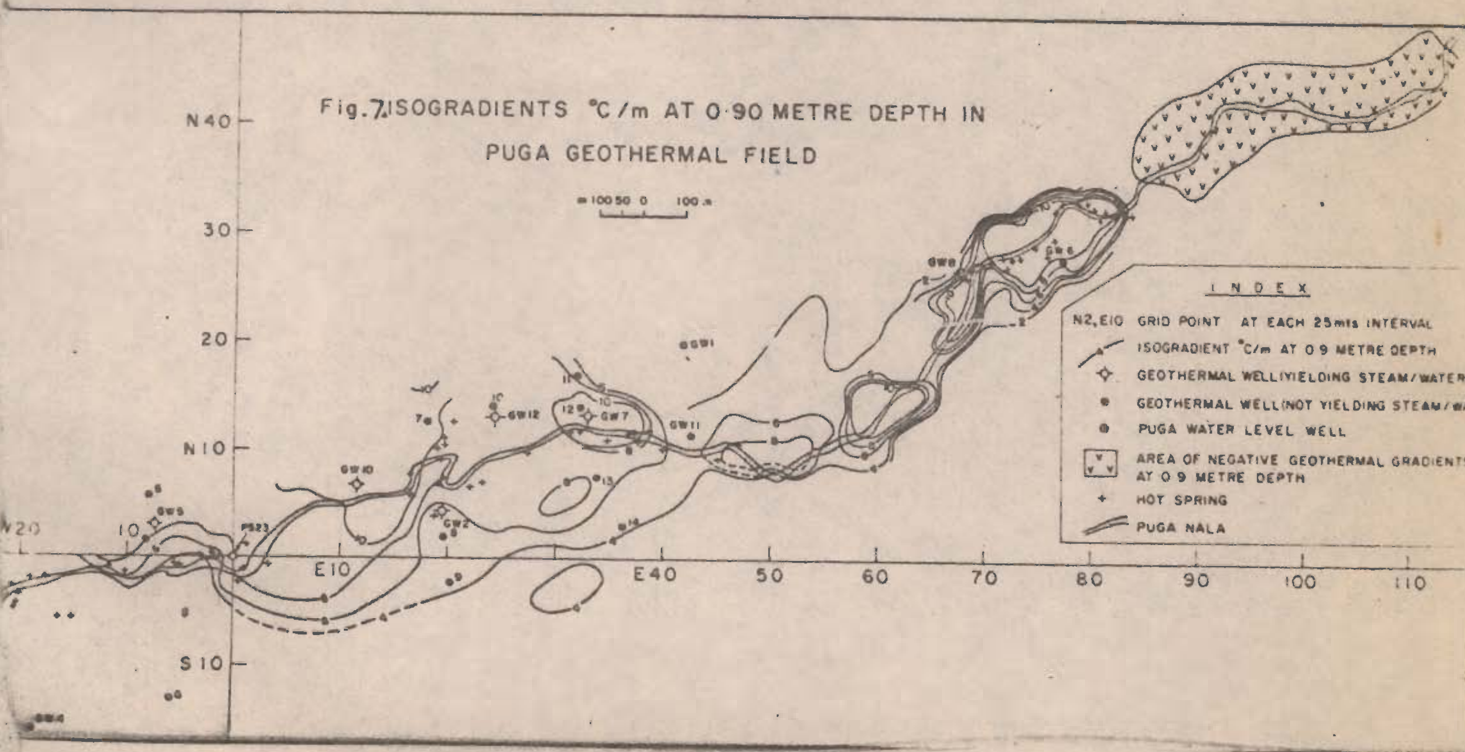


FIG. VI-6 Variation of Ground Temperature With Depth at Few Locations - Puga Geothermal Field



valley were started. Temperature observations at base stations made in the two years showed some variation but this does not affect the overall results and inference.

6.3.3. Discussions

A systematic analysis of the shallow thermal data obtained in the valley leads to the following conclusions:

i) Subsurface temperatures at 1-m depth gradually increase from the western part of the valley towards its central part (located between grid points W88 to E84, Figures VI-2,3 and 4). There is an abrupt decrease of these temperatures from about 34°C to 14°C near the grid point E84 (Figure VI-4). Towards the east of this grid point, the temperatures are still lower and this trend continues upto Samdo village (Figure VI-9).

Temperatures gradually decrease towards west of grid point W60. In fact a big low temperature zone lying in between grid points W80 and W116 observed (Figures VI-2 and 8).

ii) The maximum observed subsurface temperature at 1-m depth is 69°C (Fivure VI-4), in the vicinity of hot springs. The lowest temperature of 2°C(Fivure VI-2) was observed within the western part of the valley, suggesting the absence of feeder channels of thermal water in this part of the valley. Hot springs are also found absent here.

iii) There is an indentation of the geoisotherms between the grid points E60 and E82 (Figure VI-4), most probably, reflecting the boundaries of subsurface zones saturated with hot and cold waters and channelised flow of thermal waters. It appears that there is a sudden change in the pattern of hot water movement in the valley, near the grid point E82.

iv) Subsurface temperatures at 1-m depth between grid points W88 (W120?) and E84 (Figures VI-2, 3, 4, 8, 9 and 10) are high along both the banks of Puga nala and decreases towards the foot hills. This pattern appears to be reversed beyond E84, as per temperature profiles E86 and E106 (Figure VI-10).

v) The observed shallow temperature gradients are all positive in the surveyed area lying in between grid points W84 and E82, while these are negative in the part of the valley lying towards west of W84 and east of E82 (Figure VI-7). Two shallow drill holes (6-10 m) drilled in the western part of the valley (west of W84) also showed normal behaviour (i.e., decreasing temperatures with depth as in normal areas the annual temperature wave gets attenuated at a depth of about 20m, and only below

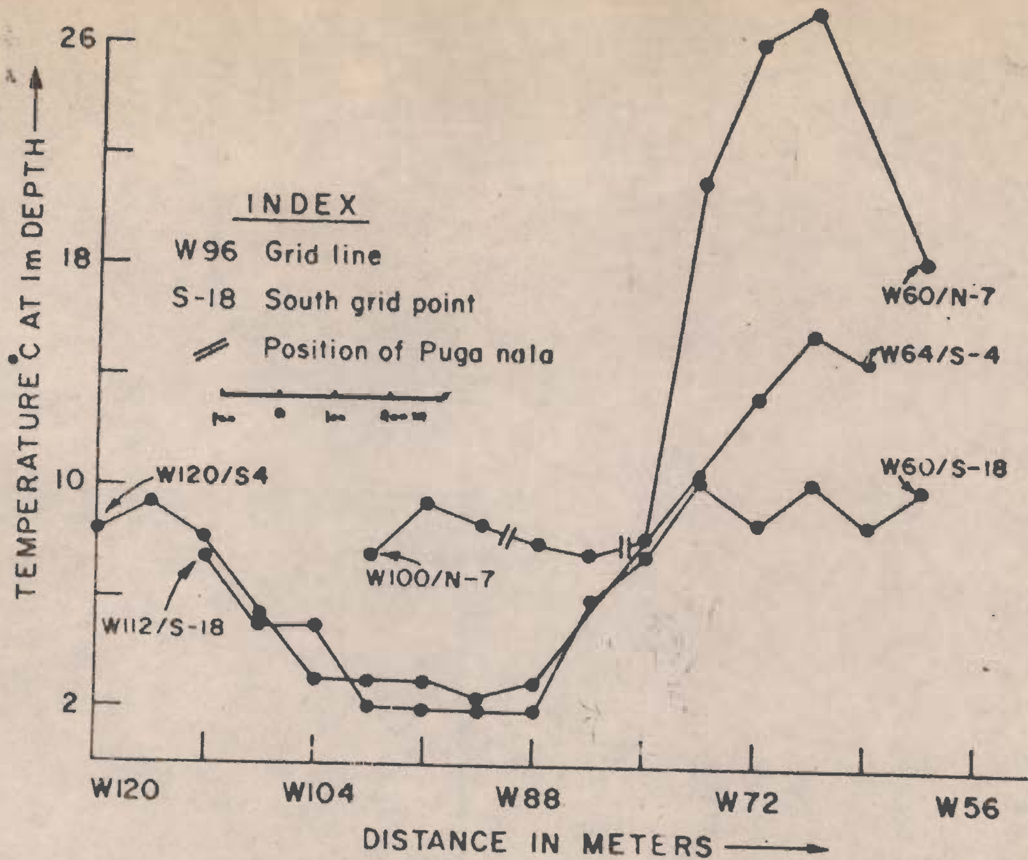


Fig 8 VARIATION OF GROUND TEMPERATURE °C AT 1-m DEPTH IN WEST—EAST DIRECTION

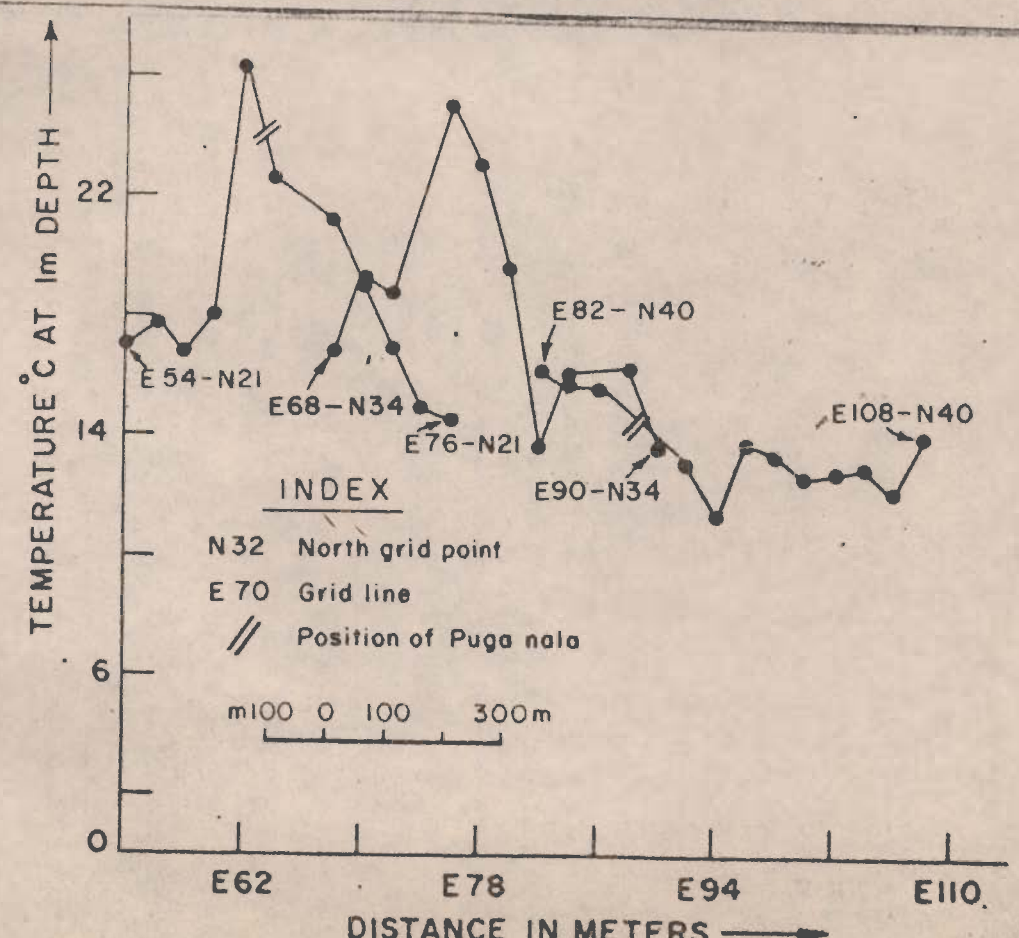


FIG. VI-9 Variation of Ground Temperature °C at 1-m Depth in North-South Direction

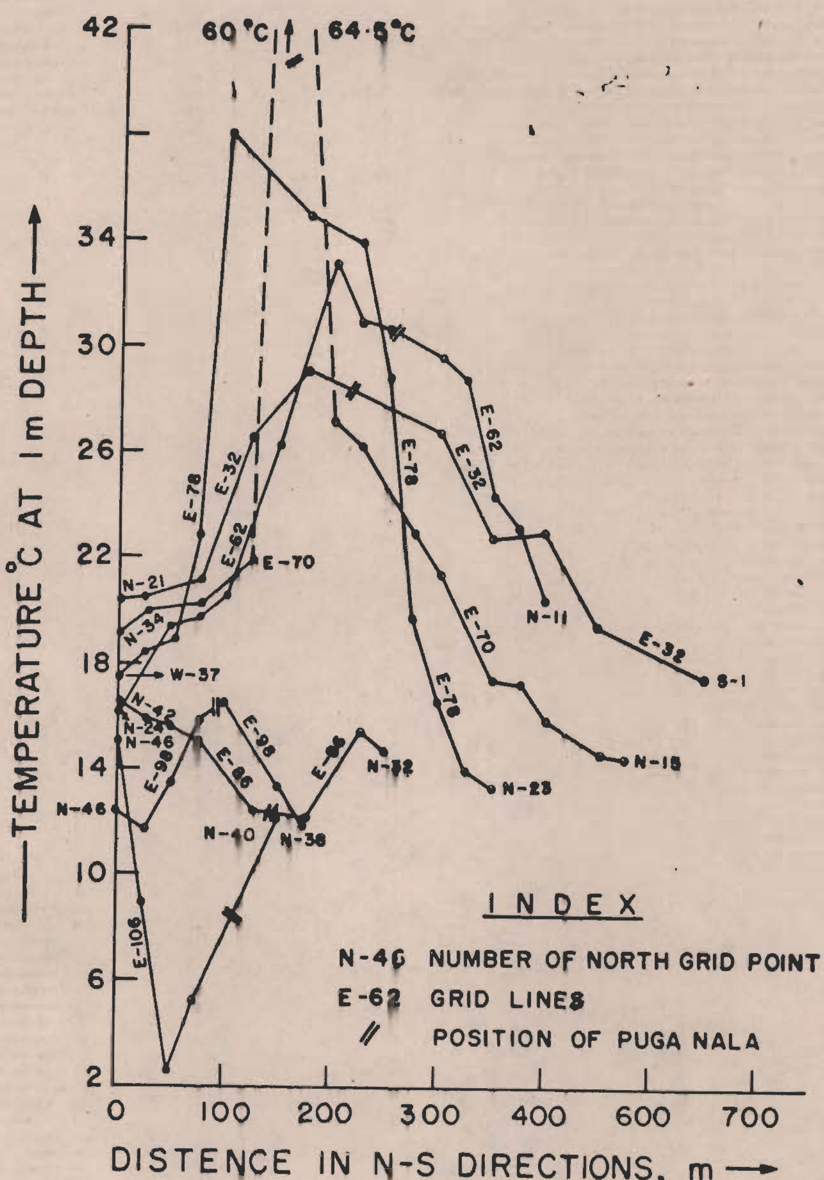


Fig.10.VARIATION OF GROUND TEMPERATURE °C AT 1m DEPTH IN N-S DIRECTION.

this depth, the subsurface temperature starts increasing). Steep lateral temperature gradients have been observed towards west of grid point W70, thus indicating either sloping basement towards west or SW or a basement fault.

vi) A 4 km stretch of the valley situated in its eastern portion between W84 and E82, encompassing an area of 3 sq.km is characterised with abnormally high subsurface temperatures and temperature gradients. This seems to have been largely caused by the upflow of hot waters and convective heat transfer at least in the upper layers. Further, the observation of negative subsurface temperature gradients in both the portions of the valley lying towards west of W84 and east of E82 strongly suggest the absence of sub-surface channels of thermal waters under those areas. In order to study the subsurface thermal conditions and to draw inferences about the location of heat source, it will be desirable to drill boreholes in these parts of the valley.

The isotherms (Figure VI-3) in the central part of the valley, based on measurements made before exploratory boreholes were drilled, gives a very good picture of the subsurface thermal conditions. The general appearance of the geoisotherms strongly suggests that a great deal of hot water and consequently much of the heat flow is transported through relatively isolated channels or group of channels, which appear to be shaped more like pipes than elongated cracks or fissures. In places, there is some suggestion or alignment of these passages along what may be more extensive faults. The trend of these passages also coincide roughly with the general course of the main stream draining it.

Boreholes 20, 20A and 20B (Figure VI-3) all were drilled for exploration of sulphur and are located at the western extremity of the sulphur mineralized area in the valley. All these boreholes encountered thermal fluids. Two more boreholes PS-8 and PS-17, which were located within 240 m towards NE of PS-20B also encountered hot water under pressure. The isotherms close up near the locations of the previous three boreholes and there is an indication that the 35°C isotherm may continue near and parallel to the location of the boreholes. This zone seems to be a good area for encountering thermal fluids of high temperature and pressure.

Borehole PS-22 (Figure VI-3) which also encountered streams of hot water is located near another 'Higher Temperature Zone'. The area under the 30°C isotherm is the largest in this zone. This seems to be a very promising zone, and warrants deeper exploration (Gupta and Rao, 1971). Geothermal wells (GW-10 and GW-12) when drilled later during 1973, 1974 in this eastern high zone encountered steam/water mixture under pressure.

Near the PS-23 borehole (Figure VI-3) the temperature contours show a NW-SE trend. This is most probably controlled either by the basement structure or by hot water flow in a narrow channel. This is another high temperature zone and borehole GW-5 when drilled later during 1975 in this zone, encountered thermal fluids under pressure, incidentally demonstrating usefulness of the rapid and quick tool for determining subsurface thermal conditions.

6.3.4 Temperature measurements in exploratory drill holes and interpretation of data

A large number of shallow holes were drilled in the valley, by the Geological Survey of India during 1969-71 for exploration of sulphur and from 1973 onwards for the exploration of geothermal resources under a joint collaborative programme.

Temperatures at various depths, in stable water column conditions have been measured in two boreholes PS-10 and PS-14 drilled for sulphur exploration and the ten other wells. The locations of all the drill holes and wells are marked in Figures VI-3 and 18 and details of comparative gradients at various depths are given in Table VI-1.

Figures VI-11 and 12 show rapid increase of temperatures with depth except in GW-6, which depict temperature reversal from 23 m downwards.

TABLE VI-1

Temperature Gradients in Shallow Boreholes at Puga Geothermal Field

Depth from Surface(m)	Depth above m.s.l. (m)	Gradient °C/m
GW-1		
8 - 23	4401-4386	2.21
23 - 27	4386-4382	1.90
27 - 31	4382-4378	1.76
35 - 40	4374-4369	2.56
40 - 45	4369-4364	1.84
45 - 50	4364-4359	1.30
50 - 55	4359-4354	0.90
55 - 60	4354-4349	0.84
GW-3		
3 - 10	4403-4396	2.55
14 - 19	4392-4387	1.94
19 - 26	4387-4380	1.36
28 - 31	4378-4375	0.62
34 - 44	4372-4362	0.32
GW-4		
7 - 12	4403-4398	1.73
13 - 16	4397-4394	2.10
16 - 20	4394-4390	0.56
GW-5		
1 - 4	4402-4399	1.05
4 - 6	4399-4397	2.83
6 - 13	4397-4390	4.60
13 - 15	4390-4388	3.28
15 - 17	4388-4386	1.65
17 - 20	4386-4383	0.00
20 - 22	4383-4381	0.55
GW-6		
5 - 10	4405-4400	1.52
10 - 17	4400-4393	0.93
17 - 23	4393-4371	0.42
GW-6 (contd)		
23 - 39	4387-4371	-0.05
39 - 41	4371-4369	-1.37
41 - 43	4369-4367	-0.59
43 - 63	4367-4347	-0.24
GW-7		
3 - 6	4397-4394	3.49
6 - 8	4394-4392	2.61
9 - 11	4391-4389	6.54
11 - 13	4389-4387	3.85
13 - 16	4387-4384	2.67
16 - 18	4384-4382	1.15
GW-8		
3 - 7	4406-4402	1.07
7 - 12	4402-4397	4.54
GW-9		
1 - 4	4404-4401	1.87
4 - 9	4401-4396	0.88
GW-10		
2 - 8	4398-4392	2.30
8 - 10	4392-4390	0.73
12 - 14	4388-4386	1.81
GW-11		
1 - 5	4397-4393	1.90
5 - 9	4393-4389	2.66
9 - 12	4389-4386	1.93
12 - 14	4386-4384	1.09
14 - 17	4384-4381	5.80
17 - 18	4381-4380	1.03

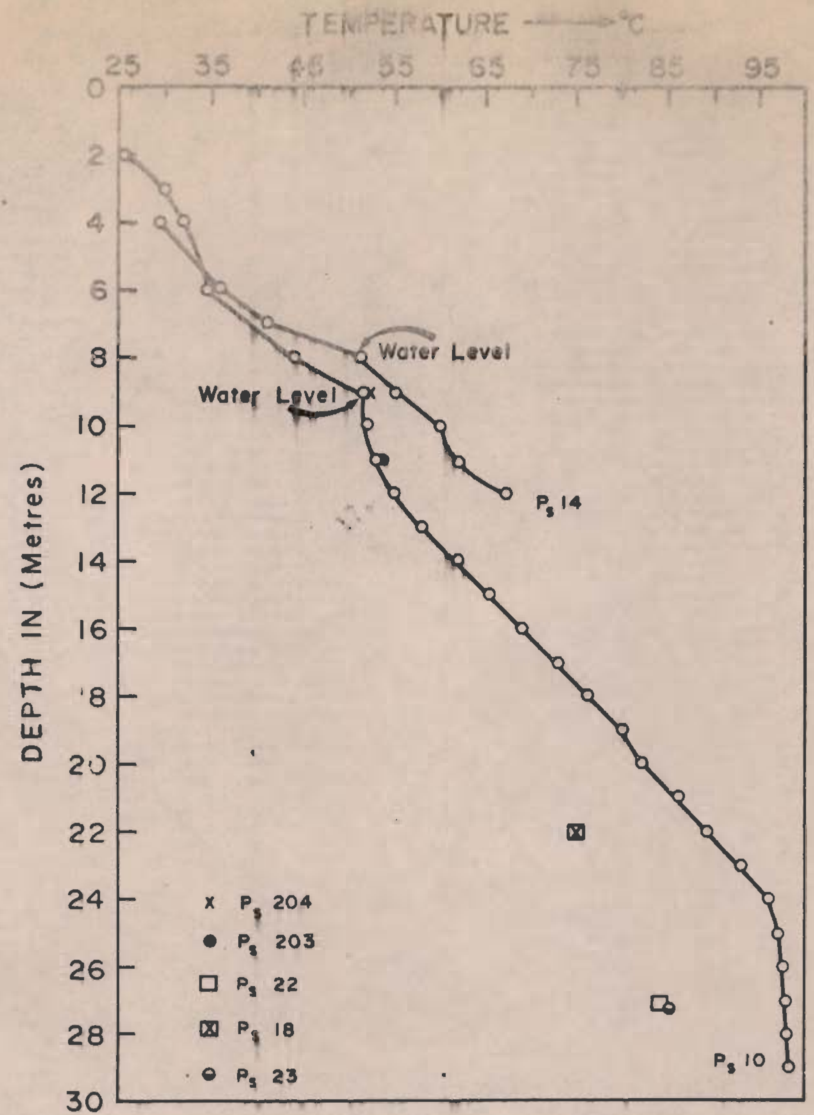


FIG. VI-11 Temperature Vs Depth Profiles of Puga Boreholes

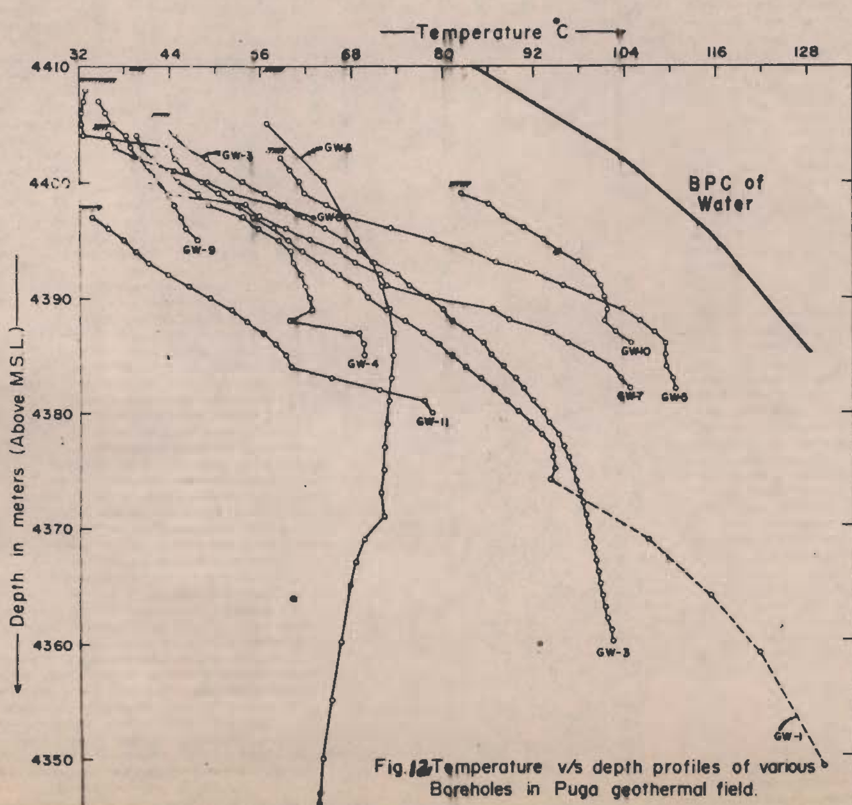


Fig.12 Temperature v/s depth profiles of various Boreholes in Puga geothermal field.

FIG. VI-13 Steam and Water Mixture Blowing from a Drill Hole Puga Geothermal Field

Table VI-1 (GW-1, GW-3, GW-5 etc.) clearly shows that the near surface layers of the central part of the valley are characterised with high temperature gradients, which generally start decreasing after a particular depth in each drill hole/well. Such a behaviour suggests that a hydrothermal convective system is located at a very shallow depth. The wells when drilled to deeper depth tapped buoyantly rising thermal fluids (Figure VI-13).

In order to have a good insight into the subsurface thermal conditions as revealed by the thermal data, let us now consider the complete system. The system under the present study is a valley filled with sediments which are generally saturated with cold waters and at places with uprising deep thermal waters containing about 1800 ppm of dissolved solids. Deep thermal waters may flow either laterally into the valley at certain depth from elsewhere or from the deep parts of the valley through connected faults and fissures. In any case, the top 20-30 m or more, as the case may be, of thick sediments can be considered as being heated through up flowing or horizontal flowing thermal waters. Additionally they may also, in some cases, encounter decending cold waters.

Temperature has a pronounced effect on the concentration of certain cations and anions in thermal waters (section 2.5.5). The ascending thermal fluids, when flowing in a top permeable layer, would naturally deposit dissolved constituents as these gradually cool near the surface. This in turn, reduces their effective porosity and renders them wholly or partially impermeable. If the thickness of the top permeable layer is appreciable, and the source temperature of the deep thermal water is 180°C, the following phenomenon must occur, in most of the cases. Depending upon the environmental conditions and the character of the thermal waters a suitable caprock will be formed at appropriate depth, thus impeding the unrestricated flow of thermal waters, which eventually would escape to the surface through certain paths, determined again by various factors. The sediments which overlie the zone of ascending thermal fluids would gradually lose their primary porosity and experience reduced permeability. Conduction and convection partly would be the mode of heat transfer, apart from mass flow, in these layers; while convection will be the dominent mechanism of upward heat flow through connected openings in the lower permeable layers.

To summarise the above arguments; in the case of Puga valley the top layers of the sediments of its central part, under which channels of thermal waters exist, have been more or less cemented and behave like a rock of low discontinuous porosity. High temperature gradients are registered in drill holes/wells in such top layers due to predominance of conduction. The lowering of the gradient values is due to the wells opening into formations in which temperature equilibrium takes place due to convective fluid motion.

In certain parts, due to saturation of near-surface layers from rapid rise of thermal waters to the surface, convection may play a significant role in heat transfer, as would appear to be the case in Puga.

The hot springs do not show intense geyser activity nor do large patches of steaming ground occur in the valley, indicating thereby the absence of boiling water conditions at shallow depths. It is also clear from Figure VI-12 that the wells registered lower temperature than the boiling temperatures for pure water for corresponding depths. All the wells are located in the thermally anomalous zone and most of them have flow, hot water along with some flash steam, thus supporting the above inference.

Reversal of temperature as registered by GW-6 appears to have been caused due to a sheet of moving water of low temperature. As regards temperature-depth data of PS-14, it is inferred that, most likely, hot water either enters the borehole or rises up along the casing from a depth of about 24 m or so and flows into the formations at a depth of about 9 m (Figure VI-11). Drill hole PS-10 also gives such an indication. The depths of water column in the boreholes was 9 and 8 m respectively.

The rapidly increasing temperatures and the high near-surface temperatures point out to a heat source nearby, in the present case, - a hydrothermal reservoir. High near-surface temperature gradients are caused by the low mean surface temperature and shallow high temperatures.

6.3.5 Determination of thermal conductivity of Puga sediments

A large number of soil samples were collected from one metre depth from the sites, where 1-m temperature gradient studies were carried out. Thermal conductivity coefficients, of seventeen samples were determined by using thermal conductivity needle probe (sections 2.1.2 and 2.2.2). The values range from 2.5 to 4.2 with a mean of 3.01. Some very high values were also obtained.

6.3.6 Measurement of natural heat loss on the surface

The heat flow on the surface, in those parts of the Puga valley where high temperatures were recorded at 1-m depth, was measured by using the heat flow transducers (section 2.4) embedded at depth of 1-m. The heat flow values obtained range from about 180 HFU to about 2,500 HFU.

6.3.7 Natural heat loss from the Puga geothermal field

Estimation of natural heat flow from a geothermal area forms a very important part of the exploration programme as it provides a basis for determining a safe and long term minimum for power production from shallow drill holes.

Conduction is the dominant mechanism of heat transfer through the soil if the temperature difference between the surface and at 1-m depth is not greater than 25°C (Dawson, 1964).

Taking the average thermal conductivity of the saturated soil as 3 m cal/cm.sec.°C, the observed temperature gradient, and the respective areas involved, the heat loss from the valley through conduction has been estimated to about 2400 kcal/sec.

It has been already mentioned (page VI-24) that the heat flow in areas which are dominated with convection, range from 180 to 2500 HFU. Taking a rather conservative figure of 300 HFU and the area as determined on the basis of 25°C geoisotherm, the convective heat loss turns out to be 1800 k cal/sec.

Heat loss through water discharge from a hydrothermal system can be best estimated through the measurement of heat content of stream/s into which spring and geysers discharge their thermal waters directly or through seepage. This is preferable to individual measurements, partly because, springs may discharge a greater proportion of their heat by thermal fluid seepage rather than through surface flow. The heat enetering into the stream due to spring discharge can be calculated by measuring the stream flow and temperature at upstream and downstream points in ageothermal area. Adopting such a procedure a value of 1300 k cal/sec was obtained.

The total natural heat flowing upwards through the process as mentioned above and lost from the surface will be the sum of the above mentioned heat losses. This is equal to 5500 kcal/sec or 20 billion cal/hr. which is equal to approximately 25 MW electric power theoretically. The heat loss from the surface calculated through equations given in section 2.5.3.4 comes out to be of this order.

6.4 Geothermal Surveys - D.C. Resistivity Soundings And Profiling

D.C. Resistivity soundings using a Schlumberger electrode configuration with a probing depth upto 0.9 km (equal to electrode spacing AB/2) and resistivity profiling with various electrode spacings along with few cross-soundings near boreholes were carried out in the valley using a NGRI resistivity metre. Electrode spread for the soundings were laid approximately parallel and perpendicular to the strike direction and were taken on both sides of the Puga nala and in the central part of the valley. The sounding curves were interpreted using a curve-matching procedure from albums of theoretical curves in conjunction with the auxilary point diagrams (Compagnie Generale de Geophysique, 1963; Orellane and Mooney in 1966).

6.4.1 Results and discussions

The value of true resistivity of various subsurface zones and the most probable thickness of the different layers, as obtained by interpretation of the field data, are given in Table VI-2. The field curves are shown in Figures VI-14 and 15.

All the soundings considered together show a large variation in the resistivity (11 to 1125 ohm.m) of the top surficial layer. However, the thickness of the top layer is usually of the order of a few metres. The wide variation in electrical resistivity of the top layer is mainly due to variation in the depth of the water table, from about one to six

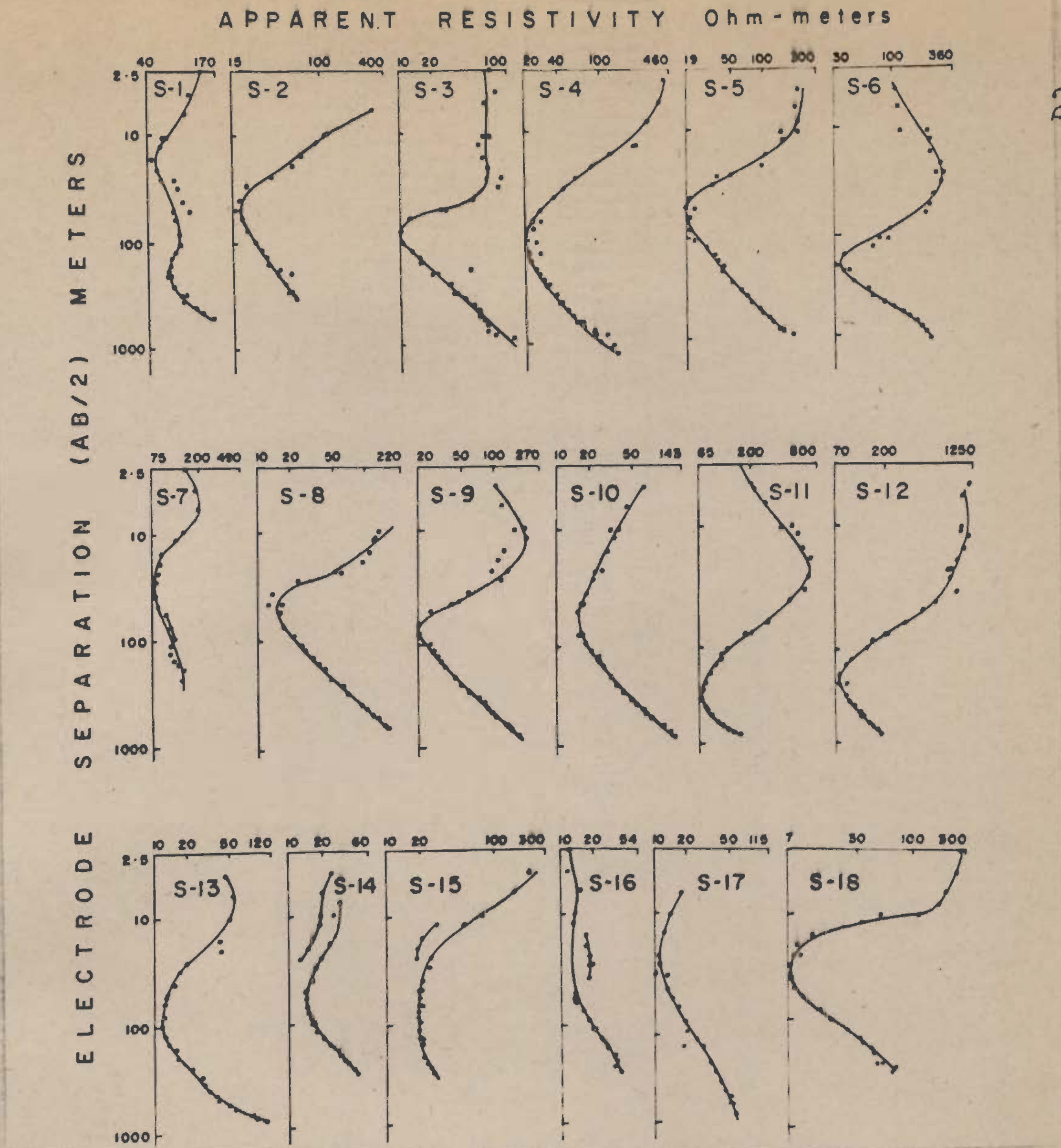


FIG. VI-14 Schlumberger Resistivity Sounding Curves
Puga Geothermal Field

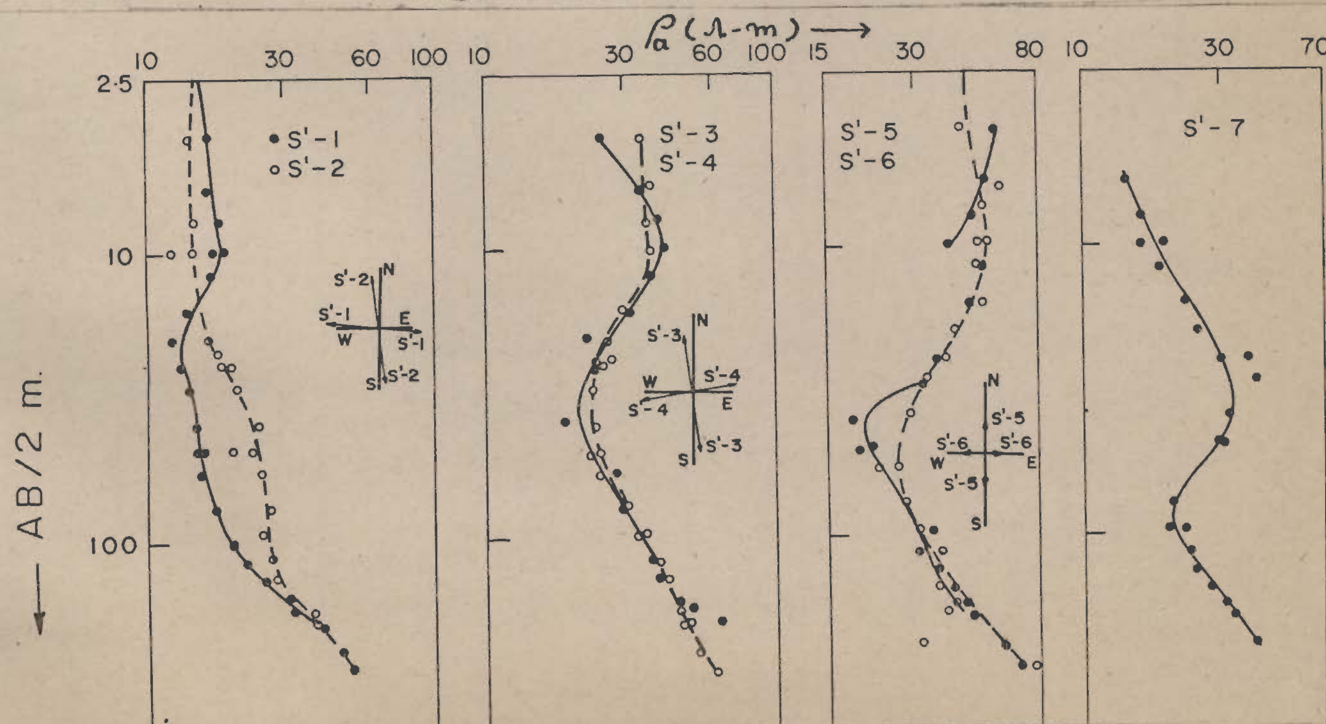


FIG. VI-15 Schlumberger Resistivity Sounding Curves
Puga Geothermal Field

metres, the presence of hot springs and spring deposits, mixing of cold surface (Puga nala) water and hot saline subsurface water, the presence of cemented breccia, and so on. The location of resistivity soundings are shown later in this chapter. The soundings are divided into the northern, middle and southern part of the valley and are discussed below.

6.4.1.1 Electrical soundings over the area north of the Puga nala

Nine vertical electrical soundings, S-1 to S-7 and S-14 (GW-9) and S-18, were carried out on the northern part of the Puga valley. Sounding S-1, in the western part of the valley near the northern flank, is located beyond the main hot spring zones. The curve of this sounding (S-1, Figure VI-15) is a typical KHK-type curve representing a five-layer section, suggesting a resistive substratum at a depth of about 180 m. Below a thin layer, thick layers of resistivity 36 to 92 ohm.m overlie the resistive basement. These are either layers of low porosity saturated with hot water or more probably layers saturated with a mixture of fresh water and saline thermal water.

Soundings S-2 to S-5 and S-18 are all located in the northern foothills. The interpretation of curves S-2 to S-5 practically provides identical results and represents a H-type behaviour. Between a top thin layer of resistivity 115 to 350 ohm.m and the resistive substratum, there exists a low resistivity zone of 7.5 to 21 ohm.m. The thickness of this zone varies from about 72 to 143 m. However, the lowest resistivity of 5.1 ohm.m has been observed at the sounding site S-18. The drilling data from various shallow boreholes (30-60 m) indicates that the subsurface layers consist of sands with grains of quartz mica and so on at the top, followed by fractured schists or breccia, consisting of fragments of gneiss, schists, phyllite quartzite, and so on, in an argillaceous, siliceous matrix. Keeping in view the electrical properties of these layers, of fresh water, of borehole and spring saline thermal waters and of the sub-surface temperature conditions, it can be safely concluded that these layers are saturated with a mixture of deep thermal water and fresh water in various proportions. It is most likely that the brecciated zone has been cemented at certain shallow depths (30 to 60 m) by the deposition of salts from the hydrothermal solution, thus making Puga geothermal field a self-sealing field. The low-resistivity layers observed by the VES form a shallow hot aquifer. However, the zones saturated fully with the deep thermal waters that ascend and flow through the hot springs are most likely of very local nature and are not extensive. This indicates the probability that either the deep thermal waters are flowing towards the valley in some channels or that they find their way up through certain fractures and fissures in the resistive basement.

Soundings S-6 and S-7 are located toward the northeast of S-5 and were taken in the area where surface traces of sulphur and borax are very infrequent and hot springs are not

TABLE VI-2

Results of electrical soundings in the Puga geothermal field

Sounding nomenclature	Location	Thickness in metres Layer 1	Layer 2	Layer 3	Layer 4	Resistivity in ohm-metres Layer 1	Layer 2	Layer 3	Layer 4	Layer 5
	Toward the north of the Puga Nala:									
S-1	W82/N10*	3.2	10.2	58.8	110.5	120.0	36.0	92.0	40.0	v.h
S-14 (GW-9)	W65/N7	6.7	82.5	-	-	46.5	13.9	v.h	-	-
S-2	W59/(N11-N10)/2	6.3	75.6	-	-	165.0	14.8	v.h	-	-
S-3	W17/(N7-N6)/2	14.5	72.5	-	-	115.0	7.5	v.h	-	-
S-4	E12/N10	5.5	143.0	-	-	350.0	21.0	v.h	-	-
S-18	(E27-E28)/2/N17	2.0	2.6	36.9	-	335.0	134.0	5.1	v.h	-
S-5	E33/(N17-N16)/2	7.9	79.5	-	-	213.0	14.9	v.h	-	-
S-6	E56/N25	2.4	21.6	158.0	-	75.0	375.0	27.36	v.h	-
S-7	E100/N45	1.9	3.2	12.78	-	140.0	280.0	32.55	167.0	-
	Towards the south of the Puga Nala:									
S-12	W100/S26	2.5	29.2	356.5	-	1125.0	900.0	63.0	v.h	-
S-11	W24/S21	2.1	14.7	472.2	-	123.0	1230.0	63.2	v.h	-
S-15 (GW-4)	W18/(S16-S15)/2	3.7	239.2	-	-	250.0	20.4	v.h	-	-
S-10	0/S11	1.6	14.8	76.5	-	140.0	28.0	9.6	v.h	-
S-9	E29/0	2.2	6.8	84.1	-	70.0	350.0	15.5	v.h	-
S-8	E52/N6	8.5	53.5	-	-	210.0	10.5	v.h	-	-
	Central part of the valley									
S-13	W38/S11	7.7	150.1	-	-	57.5	10.3	v.h	-	-
S-16 (GW-2)	E19/N4	1.9	74.2	-	-	11.0	13.7	135.0	-	-
S-17 (GW-7)	(E33-E32)/3/(N13	1.3	15.5	-	-	21.0	10.9	77.0	-	-
S'-1 (GW-5)	(W8-W7)/2/N3	2.1	3.4	2.4	71.5	15.0	18.8	3.6	21.0	380.0
S'-2 (GW-5)	Perpendicular to S'-1	1.5	16.5	7.0	131.6	21.5	14.0	52.5	31.5	101.5
S'-3 (GW-10)	(E-11-E12)/2/N7	11.0	18.1	-	-	41.5	12.5	59.5	-	-
S'-4 (GW-10)	Perpendicular to S'-3	2.5	7.5	28.0	-	31.1	38.8	14.8	90.0	-
S'-5 (PS-23)	0/0	1.7	12.6	48.2	-	70.0	45.5	14.7	100.0	-
S'-6 (PS-23)	Perpendicular S'-5	2.3	8.5	58.3	-	42.0	52.5	20.8	84.0	-
S'-7 (GW-11)	(E42-E43)/2/N12	4.8	31.2	31.5	-	12.0	42.0	10.0	190.0	-

* Grid points at 25 m interval; v.h. means very high

seen in the vicinity. The analysis of data indicates that the feeder channels of deep thermal fluids are most likely not located underneath this area. However, borehole GW-8 drilled about 100 m towards the east of sounding S-6 discharged thermal fluids at 85°C, indicating that the layers probably have low porosity. In order to have somewhat better knowledge of the subsurface underneath a nonproductive borehole, sounding S-14 was carried out with a maximum electrode separation (AB/2) of 300 m towards southwest of sounding point S-2 at drill hole No.GW-9. This sounding located on the northern flank of the valley at the extreme western end of the hotspring zone represents more or less the same behaviour as that of S-2. Lower subsurface temperatures than those near boreholes GW-7 and GW-5 were observed at 1-m depth in this area as well as in the borehole. This fact, along with the observation of low-resistivity layers of 14 ohm.m indicates that there are large lateral variations in the parameters which cause variations in the resistivity. Nevertheless, it can be inferred that the top layer of about 6 m is mostly saturated with fresh water, and below this the formation is saturated with a mixture of thermal fluids and fresh water. As mentioned above the subsurface temperatures are comparatively lower and the geothermal well did not discharge any thermal fluids, which indicates that the area lacks appropriate pressure for the flow of fluids.

The right-hand part of some VES curves (S-6, S-11, S-13) shows an inclination of more than 45 degrees, which is theoretically not possible if all four electrodes were placed in the same physical conditions. However, field experience indicates that if one of the current electrodes is on highly conducting ground and the other on resistive ground, one may get inclinations above 45 degrees for a large separation.

6.4.1.2 Soundings on the southern part of the valley

Six soundings, S-8 to S-12 and S-15, were taken in this area. Except S-8, all the soundings are located near the southern foothills. Sounding S-12 is located beyond the hot spring zone in the south-western part of the valley. The results of this sounding, when combined with those of S-1 located in the northwestern part (Table VI-2), indicate the absence of low resistivity formations in this part of the valley. However the results of VES S-11 further indicate that this behaviour continues in a more extensive area eastwards in the southern part of the valley. Under sounding locations S-12 and S-11 the resistive substratum occurs at greater depths than that at all other sounding sites in the valley. It occurs at depth of about 380 to 490 m. Above this resistive zone and under a thin resistivity layer lies a zone of resistivity of about 63 ohm.m. Without any other control it is difficult to ascribe this behaviour to any specific factor. However, it seems likely that the formations in this area are saturated mostly with fresh water.

Sounding S-15 gives a similar picture as obtained by VES curve S-4 toward the northeast, except that the thickness of the middle conducting zone is much larger here (about 240 m). Borehole GW-4 drilled to a depth of 50 m in this area has not discharged thermal fluids. Temperatures in the above borehole (Fivure VI-12) are much lower than those in the area located toward the northeast of PS-23.

The general behaviours of sounding curves S-10, S-9 and S-8 are alike, and similar to those of S-3, S-5 and S-18 soundings in the northeastern part. The location of sounding S-10 is just towards the south of PS-23, and the locations of S-9 and S-8 are towards the east. In these cases the resistive substratum occurs at more shallow depths in comparision with the areas underlying sounding points S-15, S-11 and S-12. There is an increase in the depth of the resistive substratum from 100 m to about 250 m, within a small distance of 450 m from S-10 x to S-15. From the interpretation of the curves, it appears that the resistive substratum covering a lateral distance of about 780 m between S-10 and S-9 has a more or less flat relief. A low resistivity formation of about 9.6 to 15.5 ohm.m. overlies the resistive substratum and underlies a partly saturated top layer. It also has a more less uniform thickness (76 to 84 m) under sounding points S-10 and S-9, while its thickness is appreciably lower under location S-8.

6.4.1.3 Soundings in the central part of the valley

Soundings S-13, S-16 and S-17 were taken in the flat portion of the valley located near the Puga nala. Latter the three pairs of cross soundings (S'-1 to S'-6) and sounding S-7 were taken near the producing boreholes GW-5, GW-10, PS-23 and GW-11. VES S-13 is located in the west-central part of the surveyed area and represents a three-layer section. The data indicate that a low resistivity formation (10 ohm.m) exists upto depths of 158 m in this area.

The locations of VES s-16 and S-17 were chosen in the hot zone demarcated by the 1-m temperature survey (Gupta and Rao, 1970; Gupta, Rao and Narain, 1974a). All the boreholes drilled in this area have discharged thermal fluids under pressure. The interpretation of soundings S-16 and S-17 and S'-1 to S'-7 indicates that the top layer itself consists of low resistivity formations. An intermediate layer of low resistivity 10 to 30 ohm.m(if we could combine the top two layers as single layer in some cases) overlies a high resistivity layer of about 77 to 190 ohm.m, which is interpreted as a zone dominated by vapour/gas, rather than liquid water. The top low resistivity layers most likely owe their character to condensation of steam into hot water. The rocks at these depths are likely to be fragments of gneisses and schists in an argillaceous, siliceous matrix. The maximum depth to the bottom of the vapour-gas dominated layer has not been demarcated by these soundings. The lateral extent of the vapour/gas dominated zone, as indicated by the sounding results, is about 1.25 km.

The interpreted results of VES curve S-17, which is located near GW.7 where a steam-water mixture at 125°C and a pressure of 2.5 km/cm^2 have been discharged from a depth of about 31 m, indicate the shallow nature of the vapour/gas dominated layers. However, this discrepancy may be ascribed to the assumption of horizontally stratified homogeneous layers for resistivity interpretation, which certainly is not the actual case as a large lateral contrast in electrical properties of the subsurface has been observed in the Puga valley.

The pairs of cross soundings (S'-1 to S'-6) also proved useful in avoiding a grossly erroneous interpretation which may have resulted if only a single sounding had been made at each sounding station. The curves S'-1 and S'-2 (Figure VI-15) indicate the presence of a lateral discontinuity in electrical properties in the substratum especially upto a depth of 100 m. Below this depth, both curves represent more or less the same behaviour. The striking similarity and coincidence of the results of VES curves S'-3 and S'-4 (Figure VI-15), except near the surface, demonstrate the dominance of horizontal stratification and the absence of lateral pseudoanisotropy. Sounding curves S'-5 and S'-6 (PS-23) display lateral pseudoanistropy especially near the surface. The distortions on VES S-16 (Figure VI-14), however, represent an interesting example of the effect ofstrong lateral heterogeneities on field VES curves of the Schlumberger type. This type of displacement is indicative of the placement of one of the potential electrodes over a medium of higher resistivity than that on which the center of theelectrode array is located (Deppermann, 1954; Zohdy, 1969).

The measurements in different azimuths at the same VES stations have yielded significantly different interpretations. The qualitatively interpreted result of the VES curves has been taken here to reach a somewhat geologically and geo-electrically acceptable solution. The slope of the ground in the working area is toward the Puga nala, and the results show a marked lateral inhomogenety in electrical properties of the subsurface layers. The interpretational results are constrained by these factors and in our opinion give a qualitative picture.

6.4.1.4 Resistivity profiling

In order to get a picture of the lateral variation of resistivity, horizontal resistivity profiles using Wenner electrode spacing a = 25, 50, 75 and 100 m, (towards east of E 40 grid line) and a = 100 and 150 m (from grid line W 50 to E50) were also obtained along with electrical soundings.

The isoapparent resistivity map for a = 25 and 50 m are given in Figures VI-16 and 17. The resistivity values between E46 - E62 from south to north, first decreases from the southern foot hills towards the nala and then increases towards its north. There is an abrupt increase of apparent resistivity towards the north and northeast of E56/N15 point.

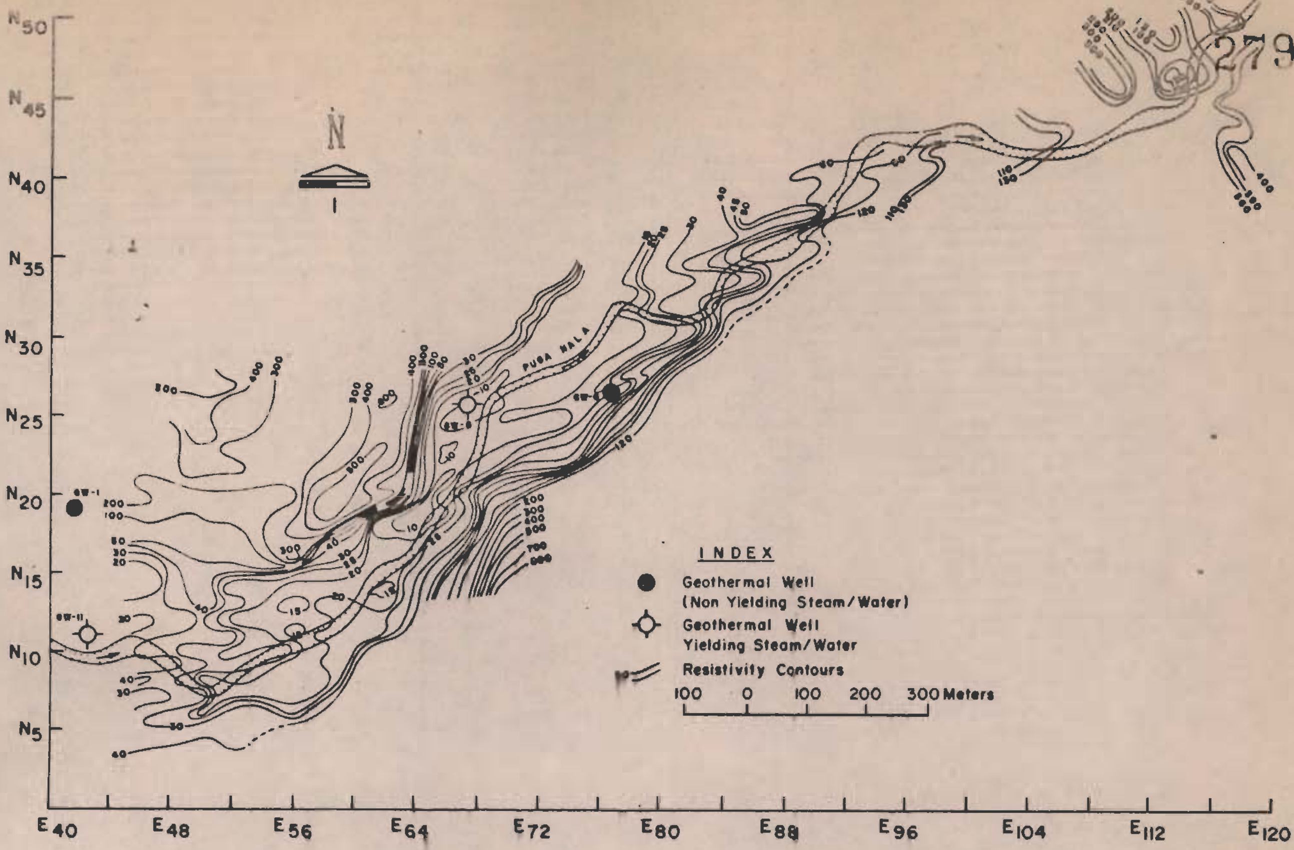


Fig. 16-ISO-APPARENT RESISTIVITY MAP FOR
WENNER ELECTRODE SPACING a = 25 mtrs. IN PUGA GEOTHERMAL FIELD

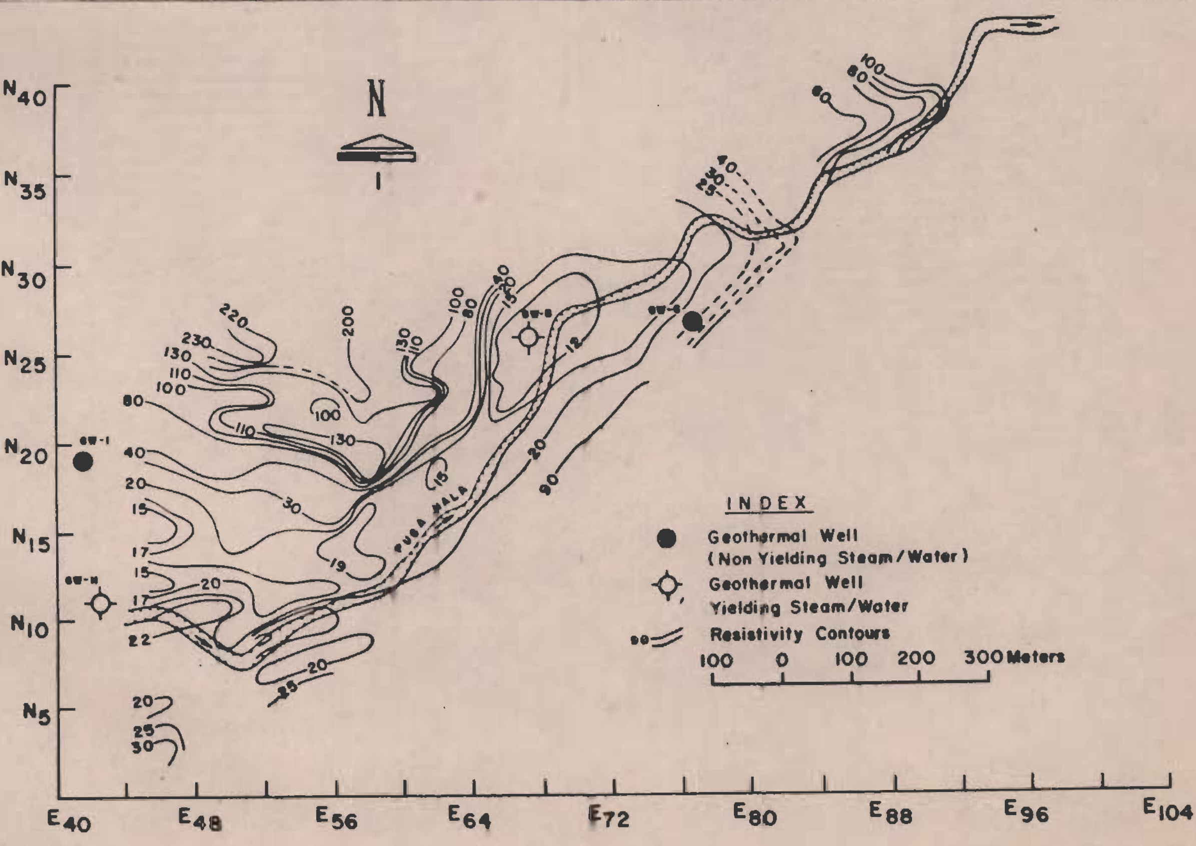


Fig. 17. ISO-APPARENT RESISTIVITY MAP FOR
WENNER · ELECTRODE SPACING a = 50 mtrs. IN PUGA GEOTHERMAL FIELD

The values increase suddenly within a short distance of about 25 m from 26 ohm.m to 340 ohm.m. However this increase in apparent resistivity shifts slightly towards north for a = 50 m (Figure VI-17) and is not so clear for a = 75 m and 100 m electrode spacings. The resistivity values along E66 almost remains uniformly low (7 to 14 ohm.m). From E70 to E84, the pattern changes and the apparent resistivity values in the northern part are lower than those in the southern part.

The results of resistivity profiling for a = 25 and 50 m gave apparent resistivity values of 10 to 20 ohm.m near the steam-water-yielding boreholes. This further supports the hypothesis that a shallow low resistivity cover exists above the layer which has been inferred to be dominated by vapour/gas.

Towards the north of the Puga nala near the grid lines E64 and E80 there ~~x~~ is an indentation of isoapparent resistivity contours (Figures VI-16 and 17). This suggests movement of hot water in a channel and separation of zones saturated with deep thermal waters and fresh water.

A traverse was also taken east of E58-N29 over this channel(?). Confirmatory data i.e., occurrence of low apparent resistivity towards the east of E66 both for a = 25 and 50 m were also obtained. Two traverses near E74-N36 point more or less parallel to Puga nala indicated that the low resistivity zone may not continue towards north.

The results of resistivity profiling for a = 100 and 150 m are given in Figures VI-18 and VI-19. The lowest apparent resistivities (13 to 25 ohm.m) with either electrode separations were found to occur near the western part of the valley. This is due to the presence of a great thickness (160 to 480 m) of saturated sediments and further supports the VES S-15 interpreted results. From the isoapparent resistivity map, it is x clear that a comparatively higher apparent resistivity (27 to 45 ohm.m) encloses that area of the valley where boreholes (GW-5, GW-10, GW-2, GW-7, GW-11 and PS-23) have discharged a steam-water mixture (Figure VI-13). The rise in the apparent resistivity is due to the effect of a highly resistive vapour/gas dominated layer. The apparent resistivities measured with an electrode seperation a = 150 m, are larger than those measured with the smaller spacing a = 100 m. This increase in apparent resistivity at larger electrode spacings also corroborates the VES data interpretation that k a highly resistive layer occurs at depth beneath a shallow low resistivity cover.

The apparent resistivity map also indicates two certain north-south features -- one located near GW-5, and another along GW-7 (Figures VI-18 and 19). The feature near GW-5 is most likely due to a change in depth of the resistive basement while a north-south hot water channel probably exists in the area in the vicinity of GW-7.

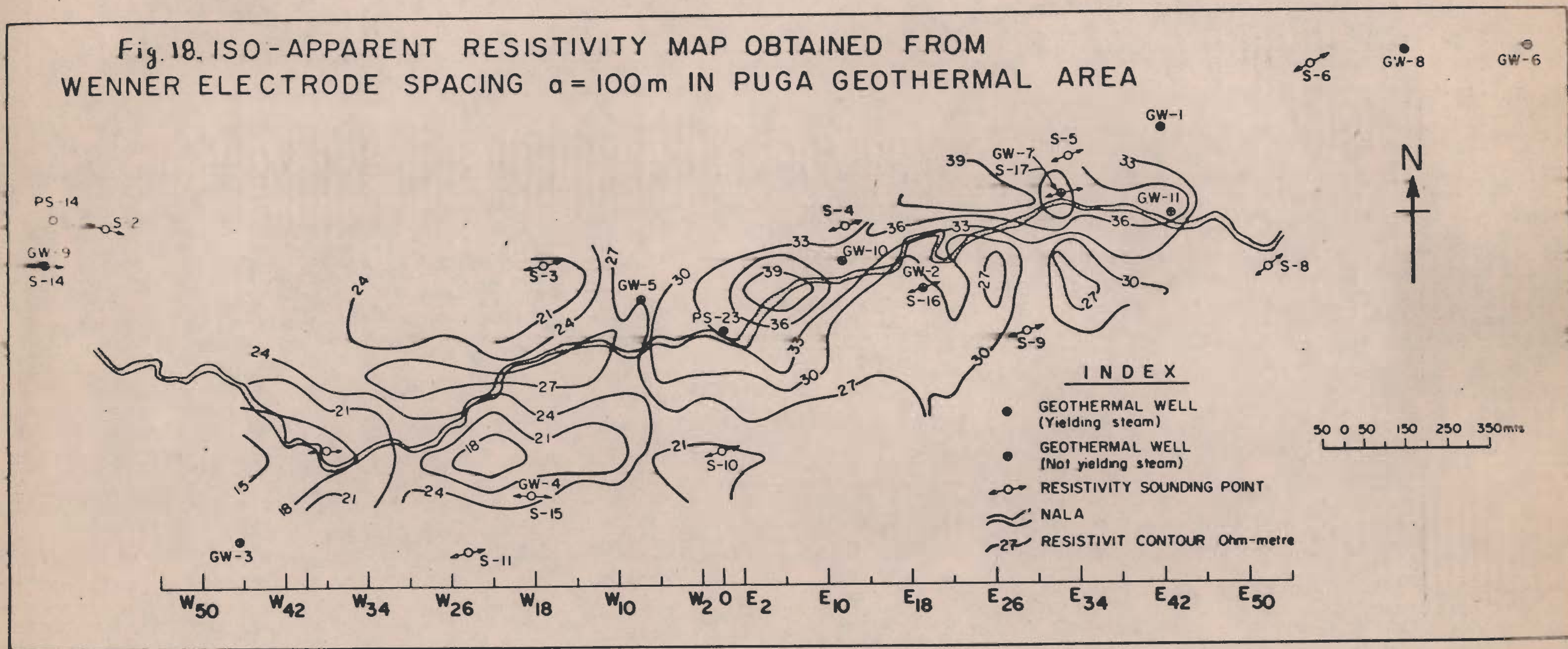


Fig. 18. ISO-APPARENT RESISTIVITY MAP OBTAINED FROM WENNER ELECTRODE SPACING a = 100m IN PUGA GEOTHERMAL AREA

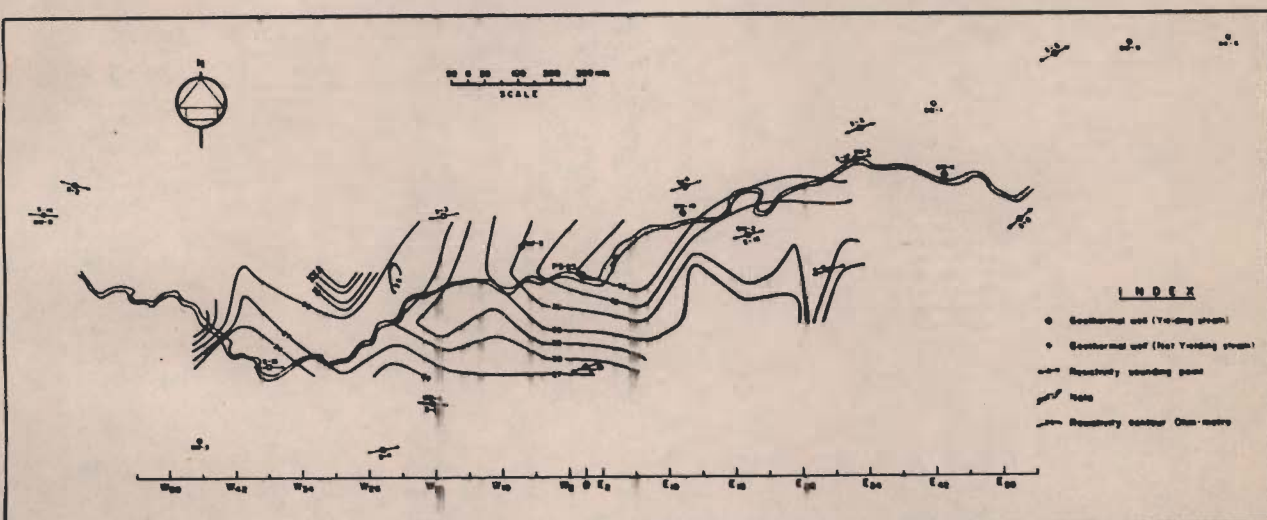


Fig. 12 ISO-APPARENT RESISTIVITY MAP OBTAINED FROM WENNER ELECTRODE SPACING a = 150m IN PUGA GEOTHERMAL AREA

6.4.1.5 Low resistivity zone

The above mentioned electrical resistivity surveys and investigations in this part of the Puga valley have delineated a zone of low resistivity (6 to 20 ohm.m) of areal extent of about 3 sq.km (4 km strike length) with some isolated zones of still lower resistivity of 3-6 oh.m. within this zone (Figure VI-20). The studies also suggest the presence of a vapour/gas dominated sub-zone of lateral extent about 1.35 km within the low resistivity zone.

A probable geoelectrical section with an area of about 1.25 k sq.km, which appears to be vapour/gas dominated is shown in Figure VI-21. Due to the limited lateral extent of horizontally stratified layers, the average values of true resistivities as obtained in a few thin geoelectric units were combined for presentation in this section. As indicated earlier, the intermediate layer, representing a resistivity of the order of 10 to 30 ohm.m is interpreted as a layer where steam condenses into hot water. A higher resistivity layer (77 to 190 ohm.m) of unknown thickness lies immediately below this layer and is interpreted to be a vapour/gas dominated layer.

6.5 Chemistry of Thermal Waters And Geothermal System

Water samples collected from hot springs, drill holes and local streams were analysed and the reservoir temperature calculated using relations given at Page II-36 to II-39.

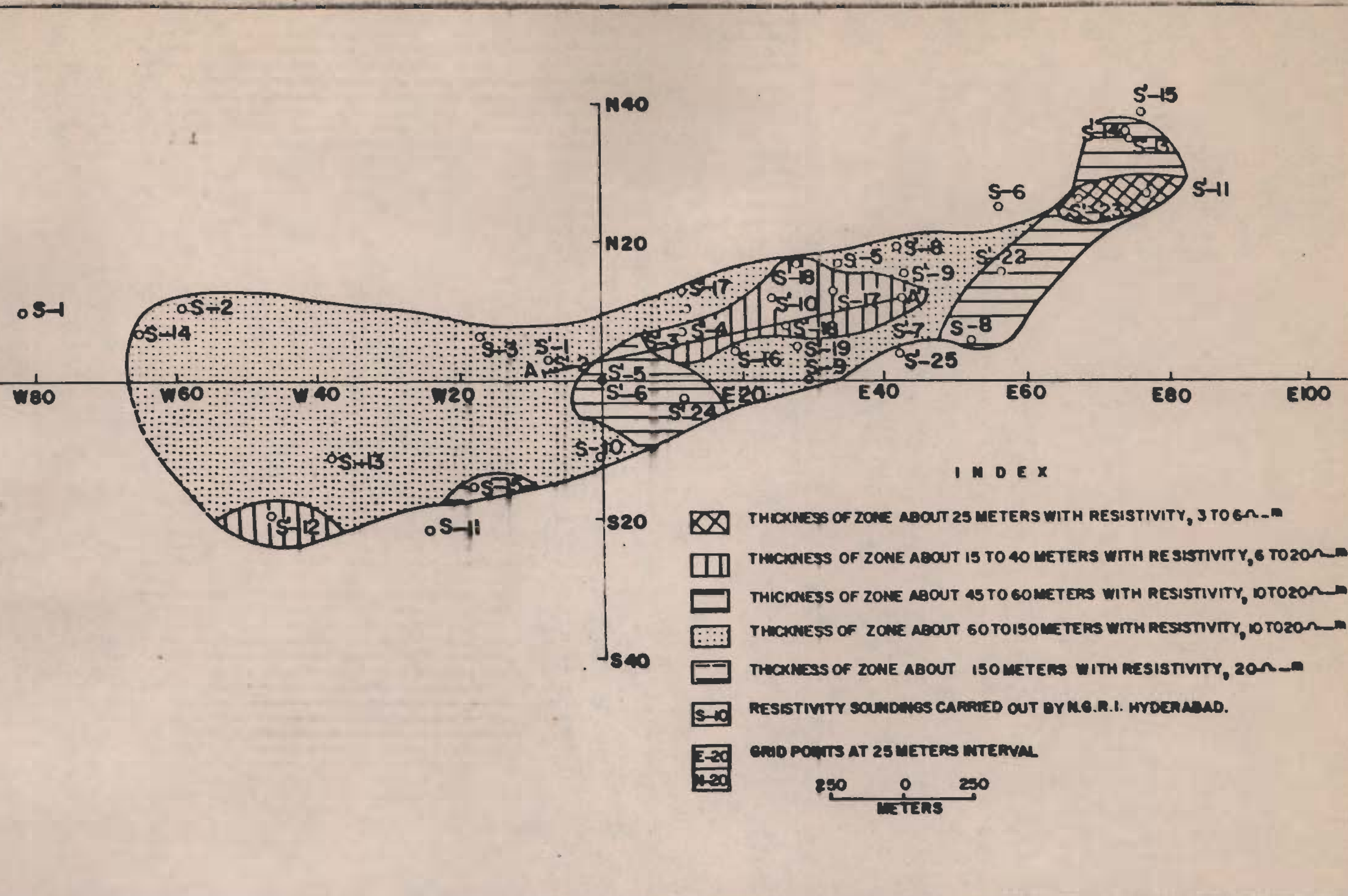
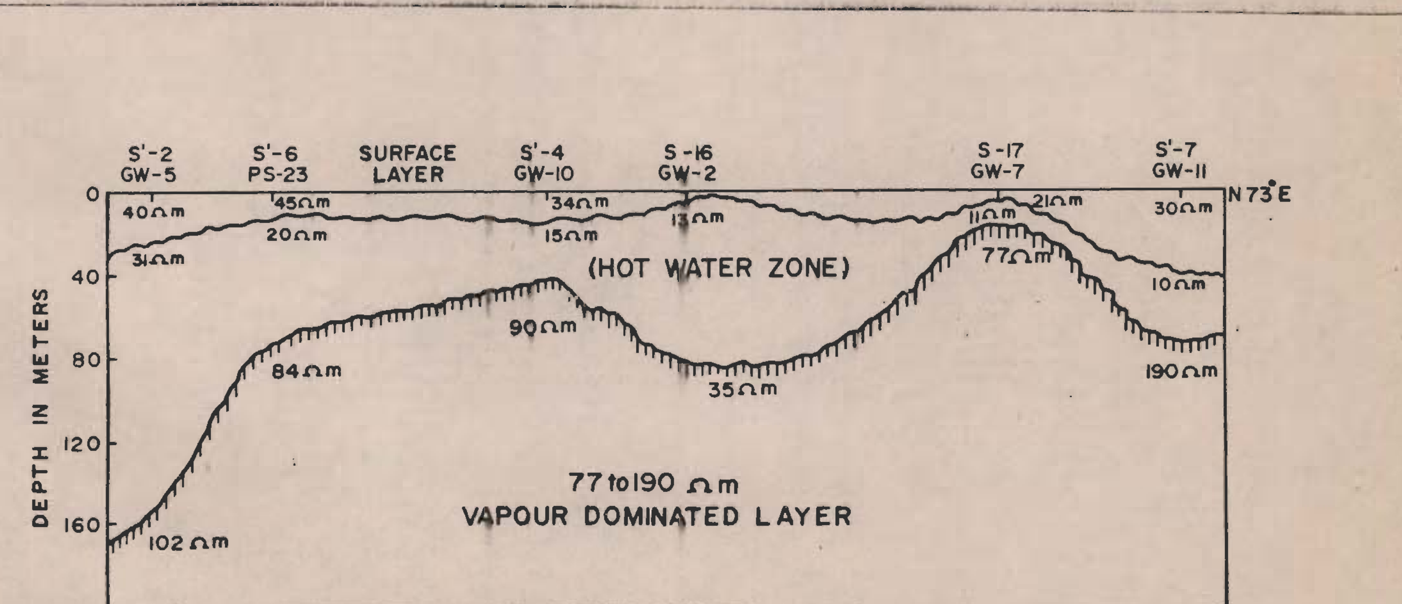


FIG. VI-20 Location of Sounding Points and Low Resistivity Zones in Puga Valley



IG. VI-21 GEOELECTRIC SECTION OF PUGA GEOTHERMAL FIELD

The results are given in Table VI-3. An analyses of the hot springs and borehole waters indicate that the total dissolved solids are of similar chemical character and concentration, indicating that the hot spring waters are not diluted at shallow depths in the valley. The probability interpretation and comparative studies show that the Puga thermal waters are somewhat similar to those of non-geyser Na-Cl-HCO_3 type, which are also associated with some magmatic components. Such waters are believed to be dominated by a relatively high F, Cl, SiO_2, B, CO_2 and Li, as well as, low Ca and Mg content. This character is more or less satisfied by Puga thermal waters which also have large (11-12 mg/lit) Cesium.

The reservoir temperatures as obtained by various geothermometers vary greatly. The temperatures as revealed by T_{SiO_2} are lower than those given by $T_{Na/K}$ and $T_{Na-Ca-K}$. Siliceous sinter deposits are observed at the surface at Puga. Occurrence of silicagel in subsurface formations was detected in drill holes, pointing to onroute precipation of silica from deep thermal waters. Silica precipation appears to have taken place largely due to conductive cooling. T_{SiO_2}, probably, would reflect the temperature of the shallow reservoir as delineated by geothermal, geoelectrical surveys and exploratory drilling. The mean T_{SiO_2} as obtained from borehole geochemical data is 175°C (Table VI-3). $T_{Na/K}$ and $T_{Na-Ca-K}$ appear to reflect the temperature of the main deep reservoir.

TABLE VI - 3

Chemical Analysis of Thermal Waters From Puga Geothermal Field

Description	PUGA BOREHOLES (30-65 m Deep)						PUGA HOT SPRINGS					
	GW-2	GW-5	GW-7	GW-8	GW-10	GW-11	S-1	S-2	S-3	PUS	PDS	PS-23
1	2	3	4	5	6	7	8	9	10	11	12	13
Nature	C&T	C&T	C&T	C&T	C&T	C&T	Clear	Clear	Clear	Clear	C&T	C&T
Smell	Sul.	Sul.	Sul.	Sul.	Sul.	Sul.	Sul.	Sul.	Sul.	-	Sul.	Sul.
pH value at 25°C	7.7	7.8	7.6	7.9	7.8	7.85	7.8	7.4	7.5	7.2	7.5	7.5
Sp.conductivity in micromho/cm	2214	2285	2142	2142	2620	2142	2213	--	2213	70	1350	2900
Total dissolve silica in ppm	200	180	180	185	182	180	210	230	200	-	112	235
Calcium	4	-	44	20	16	4	12	22	12	-	-	36
Magnesium	5	12	10	2	2	2	5	3	5	-	7	12
Sodium	780	625	512	945	520	678	525	515	755	23	400	830
Potassium	78	76	82	70	59	70	81	70	73	19	56	81
Copper	0.01	0.01	0.02	-	0.002	0.002	0.002	-	-	-	0.01	-
Zinc	0.03	0.02	0.03	0.03	0.03	0.02	-	-	-	0.05	-	0.02
Cobalt	0.03	0.04	-	-	-	0.012	-	-	-	-	-	0.032
Nickel	-	-	-	-	-	-	-	-	-	-	-	-
Stroncium	-	-	-	-	-	-	-	-	-	-	-	-
Lithium	6.5	6.3	6.3	5.9	6.3	5.7	5.9	5.5	6.2	-	3.1	6.5
Silver	0.03	0.03	0.03	0.04	0.05	0.04	-	0.01	0.01	-	0.04	0.01
TDS at 180°C	1895	1750	1818	1750	1834	1645	1750	-	1890	60	945	2592
Fluoride	18	14	14	12	14	14	16	12	14	-	7	15
Chloride	398	327	364	332	364	355	322	332	336	4	173	332
Bicarbonate	825	835	824	768	780	710	824	768	768	-	355	768
Total hardness	30	50	150	60	50	20	50	60	50	-	30	140
Sulphate	216	198	210	246	243	240	122	135	159	-	72	224

TABLE (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12	13
Weight Ratios												
K/Na	0.01	0.12	0.16	0.07	0.11	0.11	0.15	0.13	0.1	0.8	0.14	0.1
Li/Na	0.01	0.01	0.01	0.006	0.007	0.01	0.01	0.01	0.01	-	0.01	0.01
F/Cl	0.045	0.042	0.04	0.036	0.04	0.04	0.05	0.03	0.04	-	0.04	0.04
HCO_3/Cl	2.07	2.55	2.27	2.31	2.14	2.0	2.48	2.31	2.29	-	2.05	2.31
SO_4/Cl	0.54	0.60	0.59	0.74	0.73	0.73	0.39	0.39	0.47	-	0.41	0.69
(Ca+Mg)/(Na+K)	0.01	0.017	0.09	0.021	0.031	0.008	0.027	0.042	0.02	-	0.015	0.05
Na/Ca	195	-	11.7	47.2	32.5	169.5	43.7	23.4	63	-	-	23.0
Ca/Mg	0.8	-	4.4	10	8	2	2.4	7.3	2.4	-	-	3
Cl/SO_4	1.9	1.6	1.7	1.3	1.5	1.5	2.7	2.4	2.1	-	2.4	1.4
Na/K	10	8.2	6.4	13.5	8.6	9.7	6.48	7.94	1.03	1.21	7.14	1.24
T_{SiO_2}	180	173	173	175	174	173	183	189	180	-	144	191
$T_{Na/K}$	190	208	240	160	200	190	209	230	180	-	233	181
$T_{Na-Ca-K}$	217	-	181.5	177	196	212	185	199	150	-	-	135

Note: C&T = Clear and Transparent; PUS = Puga nala up stream
Sul.= Sulphurous PDS = Puga nala down stream

PS-23 = Borehole was drilled during 1970 and encountered steam water mixture under pressure. During 1973 when the water sample has been collected, the flow was similar to that of a hot spring.

The concentrations are in ppm.

6.6 Enthalpy and Reservoir Temperature

It has been estimated earlier (section 6.3.7, page VI-25) that the local natural heat transmitted from the surface of the Puga valley is around 5500 kcal/sec. The mass flow is 20 kg/sec. Assuming that the thermal fluid discharging through hot springs etc., comes ultimately from a single source and that there has been no major dilution of the hot fluid by the local ground water on the way, the mean enthalpy can be calculated as follows:

Enthalpy of Puga waters relative to 0°C

$$= \frac{\text{Total heat loss}}{\text{flow rate}}$$

$$= \frac{5{,}500 \text{ kcal/sec}}{20 \text{ kg/sec}}$$

$$= 275 \text{ cal/gm.}$$

It is difficult to estimate the total mass flow accurately but 275 cal/gm would appear to be the upper limit for the enthalpy of Puga deep thermal waters. Theoretical estimates of maximum temperature of the source for this value of enthalpy turns out to be 265°C. Under boiling-point-depth conditions, this temperature should occur at a depth of about 500 m. However the reservoir temperature as estimated from geochemical thermometers as given above vary from 183 to 240°C.

Majority of the shallow and medium depth boreholes which have been drilled in the valley tapped steam water mixture and bottom hole temperatures upto 130°C were recorded. Taking into consideration the reduction in temperature due to flashing into sream of a part of thermal water, a temperature of around 165°C has been obtained for the unflashed thermal water, which is the boiling point of water corresponding to a depth of about 65 m.

6.7 Estimation Of Power Potential

Power potential of the Puga geothermal field has been calculated using the estimated natural heat loss on the surface. The assumption made is that all the heat lost through various processes in the valley is concentrated at one place to form hot fluids. Estimate made on such a basis yield the continuous power potential.

Power potential of the shallow reservoir, delineated on the basis of various geoinvestigations, has also been estimated.

6.7.1 Continuous electrical power potential

Assuming that an energy equivalent to the estimated natural heat out put is tapped through drill holes and diverted for electricity generation, and taking 0.10 to be the efficiency of conversion to electrical energy (Nathenson and Muffler, 1975) 2.5 MW electric power can be generated and maintained from geoheat in the Puga valley. However experience in well developed geothermal fields has shown that in actual practice the withdrawl of energy is three to ten times more than the total natural energy flow from a hydrothermal area. In such a case the additional heat, which is extracted must be drawn from storage in some part of the system. This can of course be done, but for a specified time, depending on the characteristics of the system.

Banwell (1961) has given curves for estimating power produced from hot waters at various temperatures using an ideal heat engine. The engine is assumed to discharge heat into a sink (condensor) at a temperature of 30°C and to use the heat available from the water down to a temperature of 50°C. Using his curves and the estimated maximum reservoir temperature of 265°C, and assuming that all the natural heat flow is diverted to a drill hole at this temperature from the appropriate depth, the power potential will be about 275 kilowatt per kg. per sec. Multiplying this by the mass flow of 20 kg/sec gives a gross potential of 5.5 MW and, assuming an overall conversion efficiency of 60 per cent, a value of 3.3 MW is obtained for the electric power potential.

6.7.2 Power potential of the shallow reservoir

It has been mentioned above that on the basis of various geodata a 70 m thick shallow hot reservoir about 3 sq.km in area was delineated. This reservoir is likely to be of a mixed type (hot water partially vapour/gas dominated), but may be considered to be of the hot water type for power estimates.

Apart from its size, other parameters necessary to estimate the stored heat energy in a reservoir, are its porosity and enthalpy of its hot fluids.

It has been shown above (sections 6.5 and 6.6) that the temperature of the shallow reservoir is most likely to be around 165°C. Except in some narrow traps, the sub-surface temperatures in the reservoir are likely to be below the boiling point of water at that depth. Boiling of water at 65 m depth would occur at 165°C. Kavlakoglu (1975) has given a method for determining reservoir temperature or porosity based on surficial geophysical measurements. According to him the reservoir porosity ϕ^2 is given by:

$$\phi^2 = \frac{0.81 \ \rho_t (T + 21.54)}{(T + 21.54) \ \rho_r}$$

where

ϕ = porosity of reservoir

ρ_t = resistivity of the surface hot water temperature t°C

T = reservoir temperature °C

ρ_r = resistivity of reservoir in ohm.m

The resistivity of hot spring waters upto 80°C was measured in the laboratory. The porosity of the reservoir was calculated from the above relation using a value of 10 ohm.m for the reservoir resistivity; as obtained through resistivity surveys. Porosity of 21% was thus obtained for the shallow reservoir. However, for calculating the energy content, the value of porosity in case of water saturated rocks is of secondary importance. The energy content of rock and water are not very different for the same volume.

The heat stored in a hot water reservoir 'Q' is given by:

$$Q = \sigma_1 c_1 v_1 \Delta\theta + \sigma_2 c_2 v_2 \Delta\theta$$

where $_1$, c_1, v_1 and $_2$, c_2, v_2 are respectively the density, specific heat and volume of the hot water and of rock. $\Delta\theta$ is the relative temperature of hot fluid in comparision with a sink. This depends upon the use to which the heat energy is put.

Taking 2.5 gm/cc and 0.2 cal/gm as the average density and heat capacity of rocks respectively, and assuming these quantities constant throughout the depth, the total stored energy in a shallow reservoir of 3 sq.km area, for each metre of its thickness, above 10°C (considered temperature of sink for Puga valley) comes out to be 27.0 x 10^{13}cal. or 36.5 MW.

Nathenson and Muffler (1975) assume a net recovery of 25% of the stored thermal energy from a heat reservoir of 50% porosity. In the case of the first reservoir of Puga valley it is assumed that 20% of the stored heat energy can be recovered at a conversion efficiency of 0.1. Therefore, an overall recovery factor (i.e., fraction of stored energy recoverable as electrical energy) of 0.02 can be reasonably assumed.

The electrical power potential of the shallow reservoir for each metre of its thickness is thus equal to 36 x 0.02 = 0.72 MW. For a thickness of 70 m it would be equal to 50.40 MW year or equal to 2.5 MW for 20 years.

This figure is of the same order as that obtained for the continuous power potential, which has been estimated on the basis of the heat energy naturally flowing out from the reservoir. Therefore by withdrawal of such an amount of energy from the first reservoir one will only be tapping that energy which is replenished. A minimum of 2 to 3 MW of electric power can thus be safely generated for a very long period, without any serious effect, even if thermal fluids from the first reservoir of the Puga Geothermal field are tapped.

6.7.3 Total power potential

Considering a continuous generation of electric power of 5 MW, the above estimates show that full exploitation, of the shallow reservoir, would yield 3 MW from the natural heat flow and about 2.5 MW from its storage. Since a life of the order of 20 years is generally considered economically viable for a power station, these results show that the Puga hydrothermal area is adequate to meet a minimum of about 5 MW electricity requirement for a period of 20 years. However, the actual life of the hydrothermal field at this rate or even at higher rates of exploitation should be much greater. This can be estimated more accurately once the main deep reservoir is delineated.

For comparision, the natural heat out put and the estimated power potential of the Puga hydrothermal field are tabulated in Table VI-4, alongwith similar available estimates for other geothermal fields of the world.

6.7.4 Prospective use of Puga valley geothermal resources

The aforesaid estimates indicate that there is a good potential for developing geothermal resources of the valley for their multipurpose applications including power generation. Cold climatic conditions prevail during most part of the year in the Ladakh region and fossil fuels are scarce. Other potential uses of the natural hot water are for creating favourable climatic conditions for agricultural growth in hot houses, poultry farming etc. Experimental green house cultivation has already been attempted thëre. Forty one varieties of plants, including vegetables and fruits

TABLE VI-4

Comparision of Natural Heat Loss, Power Generated etc. for Some Geothermal Fields of The World.

Geothermal Field	Reservoir		Approx. area. (km^2)	Exploited area (km^2)	Power generated			Natural heat loss kcal/sec
	Temp.	Fluid			MW	Appr. area	Expl. area	
Larderello (Italy)	245	Steam	150	70	380.6	2.24	4.81	---
The Geysers, USA	245	Steam	-	6	502	-	13.70	400*
Matuskawa, Japan	230	Steam	2.1	0.4	22	9.5	50.00	---
Otake, Japan	200+	Water	5	0.15	13	2.4	80.00	1550
Wairakei, New Zealand	270	Water	16	2	192.6	12.04	96.00	100,000
Broadland, New Zealand	280	Water	10	1	66	6.6	66.00	---
Cerno Prieto, Mexico	300	Water	40	2	75	771.88	37.5	---
Puga, India	200+	Water	15	3	-	-	-	5500

(Ref: present thesis)

*Through natural discharge only

e.g., straw-berry, as well as flowers were planted and satisfactory results were obtained (Behal et al. 1976). Green house cultivation is necessary in Ladakh for the production of food stuff which are at present transported from the plains and are even air lifted during winter months when the roads are snow bound. Substantial benefits can be provided the whole year--round through supply of vegetables and tropical fruits by making a planned and coordinated effort to harness this geothermal resources. Space heating for comfortable living, defence requirements, recovery of certain dissolved constitutents etc., can also be accomplished profitably. Other potential use to which the natural steam and hot waters can be put and have been used at Puga, is for refining of borax and sulphur at the site. Commercial production of sulphur and borax is being attempted there.

6.8 Manikaran Hot Spring Area

6.8.1 Location of surface manifestations

Manikaran (Lat.32°02'N; Long. 77°25'E) is a small village situated in northwestern Himalayas in the Kulu district of Himachal Pradesh at an elevation of 1,700 m on the right bank of the Parbati river, which is a tributary of the river Beas. Hot springs occur at five different places within a distance of about 30 km in the Parbati valley viz., around the Kasol-Manikaran area, at Pulga, and Khirganga villages located about 10-15 km upstream, and at Jan located downstream of Kasol (Figure VI-22). The elevation of the valley floor changes from about 3500 (a.m.s.l) in the eastern and to about

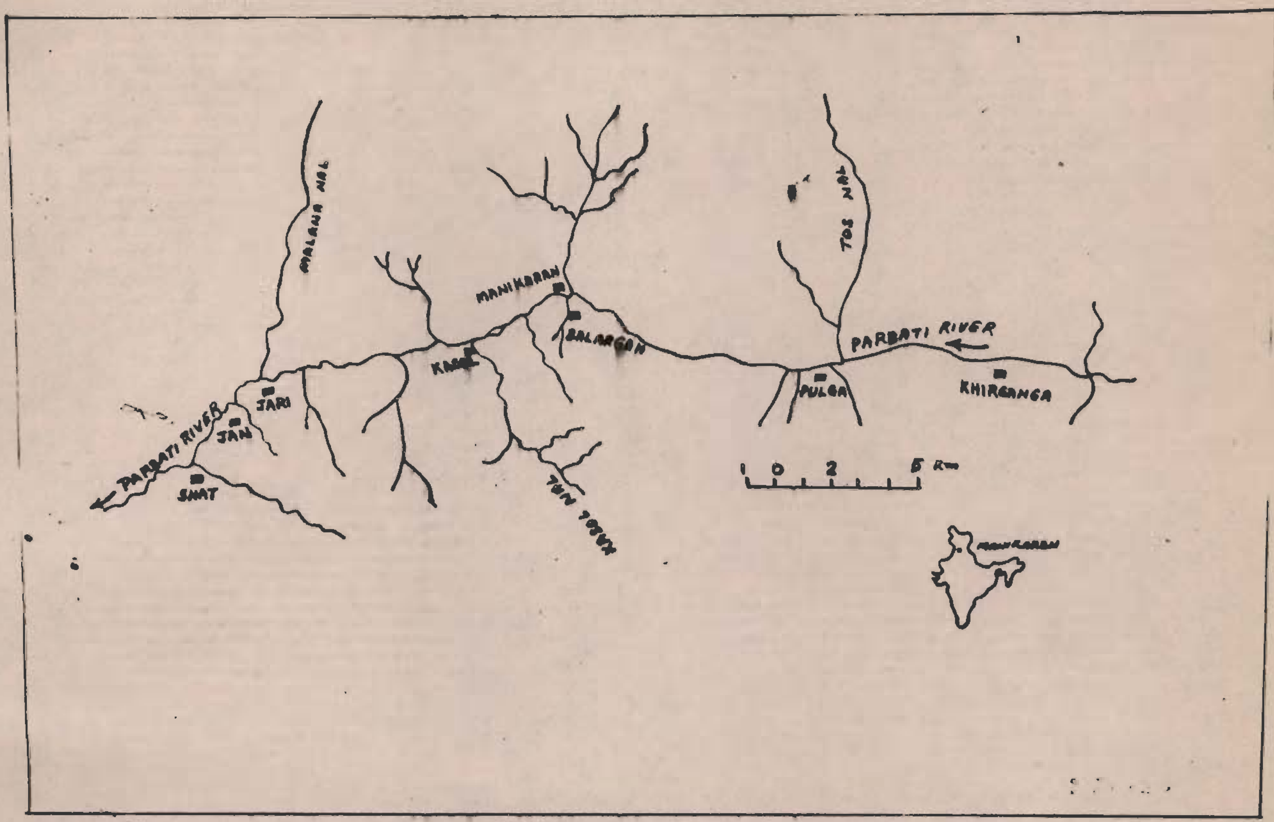


FIG. VI-22 Location Map of Parbati Valley
Hot Spring Areas

1100 m at its western end where Parbati river joins the river Beas. The majority of the hot springs, with temperature varying from 41 to 97°C, are located on the right bank of the river. The hottest springs emerge at Manikaran.

In the Manikaran area, the occurrence of hot springs, and thermal alterations of quartzite along joints and fractures are common features of the area. The hot springs are confined to several clusters, spreading along a distance of about 1.3 km beside the Parbati river (Figure VI-23). The hottest cluster is situated near the Harihar Temple which, being a pilgrimage shrine as well, attracts many tourists. The most spectacular spring in this cluster issues like a geyser near the bank of the Parbati with steam escaping for 25 to 30 seconds at intervals of two minutes. As soon as the pressure of the steam decreases, the water discharge from the spouts increases. The maximum height to which the water rises from the spouts is about 50 cm. Also, at another location near the temple, the spring water shoots to about a metre at an angle of about 45°. There are two more locations on the Manikaran terraces where water spouts and steam escape under pressure. Many other comparatively lower-temperature springs emerging from the river terrace deposit, which overlies quartzite, are also located in the Manikaran area (Gupta, 1974). The thickness of the terrace deposits as estimated from electrical soundings is about 25 m (Gupta et al. 1973). As the quartzite is blanketed by the river deposits, most of the springs on the terrace emerge through these deposits. Hot springs issuing from the

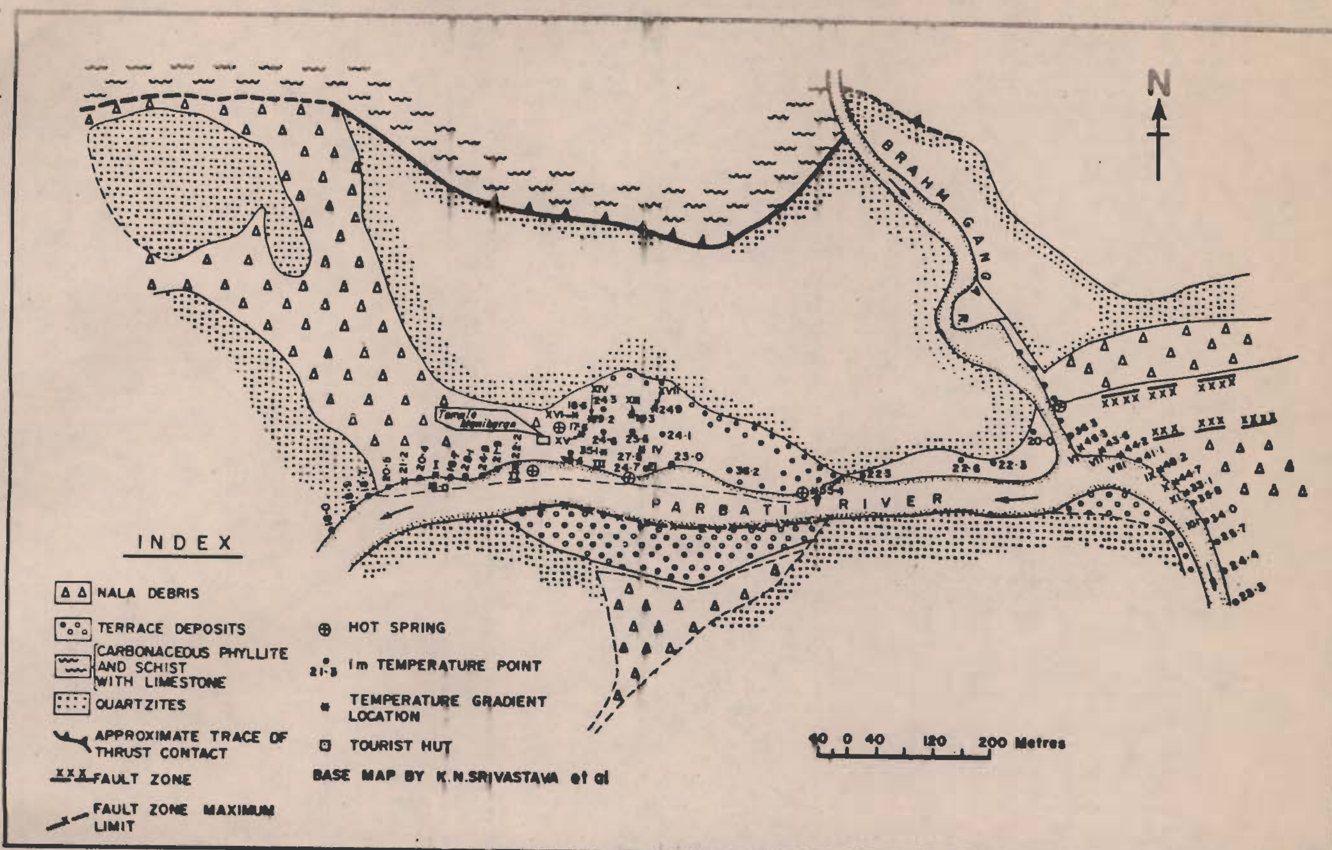


FIG. VI-23 Geological Map - Manikaran Hot Spring Area joint planes in the quartzites are seen at other locations. One cluster of springs with a temperature range from 41 to 55°C is also situated along the Brahmaganga river near its confluence with the Parbati river. Practically all the springs emit CO_2 and H_2S. Two hot springs at Jan, twelve at Kasol and three at Pulga have been noticed manifesting. However their temperature and discharge are much lower than at Manikaran.

6.8.2 Geological setting

The Parbati valley geothermal field lies to the southwest of the Central Crystalline Axis of the Himalaya where profuse igneous activity took place during Tertiary period (Pascoe, 1973).

The rock types found in the Parbati valley region consists of quartzites, gneisses, gneissic granites, schists, phyllites as well as some traps and basic intrusives.

The main formation in the Manikaran-Kasol region is a thick sequence of a massive, well jointed, white to greyish and pinkish quartzite with minor phyllite, schists and lenticular bands of limestone. The latter formations are seen occurring within the Manikaran quartzite between Jeori and Jan. Manikaran quartzite is exposed between Manikaran and Jari, and has a regional dip in a north-east direction ranging from 30° to 50°. Towards the south of Manikaran and Jari is exposed a granitic body of batholithic dimensions covering an area of about 500 sq.km. This granite has been mapped and named as Bandal granite by Sharma (1977) who inferred it to be a magmatic intrusion into the quartzite and equivalent to Panjal Traps i.e., of post-Carboniferous age. The age of Manikaran quartzite according to Jangi et al. (1976) is Devonian, while Bhanot et al. (1976) report a 1220 m.y. Rb/Sr whole rock isochron age of Bandal granite.

The Manikaran quartzite is highly jointed and exhibits three planes of joints. These joints hold great significance as for as the movements of hot waters at shallow depths at Manikaran, and Kasol is concerned. The Parbati valley has cut a deep gorge for a considerable length approximately normal to the strike of the Manikaran formations (Quartzites with subordinate phyllite and schists).

The next important rock formation is a group of basic schists and green phyllites, trap and bands of quartzite, probably of volcanic origin during Pre-Cambrian. The Manikaran quartzite is overlain by the gneisses, schists and carbonaceous phyllites of the Kulu formation. The contact between Manikaran and Kulu formations appear to be a major thrust contact and probably forms a part of Central Himalayan Thrust, which is an extensive feature related to Cenozoic folding in the Himalayan mobile belt. Kulu formations are thrust over the Manikaran formations and are older than the quartzite (Pre-Cambrian). Hot springs of Pulga and Khirganga emerge through these overthrusted sheets, which in the upper reaches of Parbati valley overlie the Pre-Cambrian granites and gneisses of the Central Himalaya.

6.8.2.1 Hydrogeological conditions

There are two distinct hydrogeological units in the valley. At Manikaran-Kasol area the first unit comprises of valley fill material and alluvial cones etc. Thickness of this unit upto a depth of 30 m has been revealed by D.C. Resistivity surveys (Gupta et al. 1973). It contains water under unconfined conditions. This unit is underlain by the Manikaran quartzite (which is very thick, highly jointed and fractured) and associated rock formations. The second unit has interconnected joints and fissures and forms the main potential reservoir for hot water and allows deep circulation of meteoric water. Melting of snow at upper reaches of the valley and the rainfall contribute to the recharge of both the aquifers. Rivers and streams may also add to recharge but it is likely that their contributions is more towards dilution of deep thermal waters.

6.8.3 Hot spring and river water chemistry

Water samples from five Manikaran hot springs and the Parbati river were collected and analysed for their chemical constituents. The chemical analysis data and the weight ratios of important cations and anions are given in Tables VI-5 and VI-6 respectively. The waters are of a mildly alkaline nature of pH value 7.3 to 7.6, possibly controlled by HCO_3^- reaction. CO_2 was detected in the gases bubbling out of the springs.

TABLE VI-5

Chemical analysis of Manikaran hot spring and Parbati River waters, in mg/litre

Description	Manikaran hot springs (1)	(3)	(4)	(5)	Parbati River
Temp. °C	97.0	75-92	81.0	45.0	13.0
pH at 30°C	7.3	7.4	7.35	7.6	7.2
Sp.cond.at 25°C in micromhos/cm	878.2	871.1	892.5	606.9	47.0
TDS	555.0	609.0	510.0	390.0	33.0
Li	1.1	1.0	0.8	0.6	ND
Na	94.0	76.0	88.0	53.0	7.0
K	21.0	19.0	18.0	15.0	3.0
Ca	36.0	32.0	32.0	30.0	6.0
Mg	10.0	10.0	11.0	10.0	3.0
Fe	0.04	0.05	0.05	0.07	ND
Co	0.014	ND	ND	0.018	ND
Cu	0.019	0.028	0.0095	ND	ND
Sr	0.135	0.135	0.140	0.133	ND
Ni	0.0018	0.0023	0.0009	0.0018	ND
Ag	0.01	0.01	0.01	0.01	ND
Zn	0.12	0.125	0.127	0.121	0.122
B	1.0	1.5	2.0	1.0	ND
SiO_2	120.0	115.0	120.0	110.0	ND
F	1.0	0.5	0.5	1.0	0.5
Cl	123.0	100.0	104.0	64.0	1.0
SO_4	45.0	48.0	45.0	46.0	ND
HCO_3	177.0	188.0	175.0	177.5	58.0

ND means not determined

The analysis shows that there is a predominance of Na^{+}, Ca^{++}, HCO_3^{-}, and Cl^{-} ions in the spring waters while Mg^{++} is present in moderate concentrations, NaCl and $NaHCO_3$ being the dominant groups. The concentration of dominant cations and anions indicates that the spring waters are of Na-Ca-HCO_3-Cl type. The dissolved solids (TDS) are present only in low concentrations and vary from 390-609 mg/litre. The presence of SO_4 ions could be due to the interaction of deep thermal water with granites and gneisses containing traces of sulphur (Barth, 1950).

Metallic trace elements have also been quantitatively determined in the spring waters by use of an atomic absorption spectrophotometer. Minor quantitites of the elements Co, Cu, Sr, Ni, Ag and Zn have been detected and their quantities are given in Table VI-5.

Chemical geothermometers (Section 2.5.5, page II-70), a recently published Magnesium corrected Na-K-Ca geothermometer (Fournier and Potter (1979) and a new K-Mg-Ca geothermometer, developed by Giggenbach (Personal communication, 1980), have been used for the estimation of temperature in the geothermal reservoir for the Manikaran area. The results are given in Table VI-7.

6.8.4 Mixing of thermal and cold waters

The appreciable difference between the subsurface reservoir temperatures, as inferred by geochemical thermometry,

TABLE VI-6

Weight ratios of the important constituents in Manikaran hot spring waters

Description	Hot Springs 1	3	4	5
K : Na	0.22	0.25	0.20	0.28
Li : Na	0.012	0.013	0.011	0.011
F : Cl	0.008	0.005	0.005	0.016
B : Cl	0.008	0.015	0.019	0.016
HCO_3 : Cl	1.44	1.88	1.68	2.76
SO_4 : Cl	0.36	0.48	0.43	0.72
(Ca + Mg) : (Na + K)	0.40	0.44	0.41	0.59
Na : Ca	2.61	2.37	2.75	1.77
Ca : Mg	3.60	3.20	2.99	3.00
Cl : SO_4	2.73	2.08	2.31	1.39

TABLE VI-7

Estimated reservoir temperature of Manikaran Geothermal Field

Spring No.	Surface Temp.	$T_{Na-K-Ca}$	$T_{Na-K-Ca}$ mg correction.	$T_{K-Ca-Mg}$*	T_{SiO_2}
1	97	210	-	-	-
2	75.9	201	121	113	145
3	81	204	114	102	148
4	45	215	-	-	143

Temperature in °C

* $T_{K-Ca-Mg}(°C) = 43.2\,A + 46.7$

$A = \log_{10} K^2/MgCa$

Values are in ppm (Giggenbach, unpublished, personal communication) and the observed temperatures of the hot springs of an area having high discharges can be explained by the mixing of deep water with shallow cold water, as happen at Manikaran where the temperature and chloride content of the hot springs show large variations. The fraction of shallow cold water mixing with deep thermal water has been estimated independently from such parameters as temperatures of hot springs, estimated reservoir temperature, and mean annual temperature of the area, and has also been estimated from the combined use of enthalpy, measured dissolved silica of hot springs and surface cold waters, and the data on the solubility of quartz at elevated temperatures as given by Morey, Fournier and Rowe (1962).

Considering the average annual temperature of the Manikaran area to be 10°C, the average temperature of non-boiling springs to be 80 to 90°C, and the reservoir temperature to be 195°C (as estimated by Na-K-Ca geothermometry), the fraction cold water mixing with deep thermal water comes out to be 0.62 to 0.57. The mixing fraction can also be calculated from the following relationships using values of enthalpy, the solubility of quartz at various temperatures, and the temperature and SiO_2 content of nonboiling springs.

$$E_{Spring} = E_{hot} X + E_{cold} (1 - X) \quad .. \quad (1)$$

$$Si_{Spring} = Si_{hot} X + Si_{cold} (1 - X) \quad .. \quad (2)$$

Where,

E = Enthalpy of water.

Si= Silica content of spring water, deep hot water and surface cold waters.

X = Fraction of deep thermal water present in the spring water.

Solving for X From equations (1) and (2)

$$X = \frac{E_{spring} - E_{cold}}{E_{hot} - E_{cold}} \qquad .. \ (3)$$

$$X = \frac{Si_{spring} - Si_{cold}}{Si_{hot} - Si_{cold}} \qquad .. \ (4)$$

Enthalpy, or total heat content, and the concentration of silica in water are functions of temperature. Using these functions and equations (3) and (4), values of X and temperature were calculated earlier. However, Fournier and Truesdel (1974) suggest a simple graphic procedure. In this graphic procedure, using equations (3) and (4), the values of X are calculated for a set of enthalpies at different temperatures and also for a set of solubilities of quartz at various temperatures upto 300°C. The point of intersection of the curves representing the above mentioned variations provides the fraction of deep thermal water present in the spring waters and the probable reservoir temperature. The estimated fraction of deep hot water present in the spring waters obtained by this method turns out to be 0.35 - 0.46, with an average value of 0.4 for thethree springs S_2, S_3 and S_4 of the Manikaran area. An overall average value of 0.60 would

would be in agreement with that found by temperature considerations only. The spring S_1 in which boiling occurs, has not been considered. Also the mixing fraction of the spring S_5 has not been estimated as the spring emerges from sand and gravels and may have picked up extra silica. The aquifer temperature obtained by this method varies from 190 to 210°C with a mean of about 202°C.

6.8.5 Tritium measurements

A few water samples representing hot springs and surface water were analysed for their tritium content. The water samples were converted to methane gas, and tritium activity was measured using an Oeschger gas proportional counter. Table VI-8 gives the result of tritium analysis, and the values are reported in tritium units. One tritium unit (TU) is defined as one atom of tritium per 10^{18} atoms of Hydrogen. The tritium contents of hot springs are found

TABLE VI-8

Concentration of Tritium and Chloride in Waters of The Manikaran Area.

Description	Temperature. (°C)	Cl mg/lit	Tritium concentration (TU)
Manikaran hot spring			
Cluster No. S-1	97	123	14
Cluster No. S-3	75-92	100	37
Cluster No. S-4	81	104	58
Cluster No. S-5	45	64	56
Parbati river water	13	1	103
Cold water from the hill	11	-	173

to vary from 14 TU to 58 TU as against concentrations of 103 TU and 173 TU for the surface streams. The value of 173 TU corresponds to snowmelt water originating at higher altitudes, probably from precipitation received 1963-64, which had a peak tritium concentration. The concentration of 103 TU is typed of the Parbati river in general.

The concentration of tritium in spring water is higher than that of the pre-bomb tritium levels for 30°N (5 to 10 TU), which could be due to shallow circulation with a short turnover time before the meteoric water emerges as spring water, or to the mixing of deep thermal waters with surface waters. It has been shown above that the Manikaran spring waters are of a mixed type, that is deep thermal waters emerging from joints and fissures become mixed with cold waters at shallow depths. The variation of the tritium concentrations of various springs (Table VI-7) also indicates the latter possibility. There is an inverse relationship between the tritium contents and the temperatures of springs. This indicates a lesser degree of mixing for the springs having higher temperatures. The chloride contents of various spring waters also show a similar behaviour. The mixing model, taking the general mixing fraction of shallow cold water with deep thermal water to be 0.6, indicates no tritium in deep thermal waters, thus pointing out that the deep thermal waters in the hot aquifer have a long residence time and therefore occupy a large subsurface reservoir, or are slow moving; Long periods of time ($>$20 years) can easily be required for the meteoric water to percolate to depths before flowing out. The tritium contents of drill-hole waters measured in another potential geothermal field at Puga (Gupta and Sukhija, 1974) also show dead water (tritium-~~ferr~~ free) at depth.

6.8.6 Natural heat discharge

The total surface discharge of thermal water from the hot springs of Parbati valley is around 40-50 lits/sec. However, appreciable seepage occurs in the river, and the total discharge is not likely to be lower than 100 lit/sec. and may be even around 150 lit/sec. Accurate figures of discharge can be obtained from an analysis of temperature, flow rate and the chlorine content values of the water of Parbati river measured at a number of selected locations. The maximum discharge occurs at Manikaran where the temperature is also maximum. Considering the discharge and temperature of various hot springs centres of Manikaran, an average temperature of about 75°C has been estimated, and heat loss of 7500 cal/sec or even more.

6.8.7 The Manikaran hydrothermal system

In the Parbati valley, hot springs are found to occur in a number of villages adjoining Manikaran. These are, apart from Manikaran itself, Jan, Kasol, Pulga and Khirganga all within a distance of about 30 km (Figure VI-22). A synthesis of various ~~ge~~hydrogeological and shallow thermal studies, geoelectrical and seismic surveys, geochemical and isotope studies, micro seismic investigations and exploratory drilling (Gupta et al. 1973; Jangi et al. 1976; Giggenbach, 1977; McNitt, 1977; Chaudhary, 1979; Kumar with Gupta and other, 1980), leads to the following conclusions regarding the springs of the Manikaran system.

i) That the thermal waters so far tapped through drill holes in the Parbati valley appear to have been diluted, maximum dilution occurring at Kasol, followed by Manikaran and Khirganga.

ii) That hydrogeological studies alongwith oxygen and hydrogen isotope measurements show that thermal waters originate at a higher altitude than those of spring sites. This implies that snow melt in the higher reaches must contribute substantially to the charging of the deep aquifer, apart from rainfall and that the circulation is probably quite deep facilitated by neotectonic faults in the region.

iii) That although no dated or geologically inferred recent igneous intrusions have been found to occur around Parbati valley, there indeed might be such bodies emplaced not too deep below the surface in the wake of orogenic activity.

iv) That whilst near surface temperature studies and borehole temperatures in the Manikaran-Kasol sector are consistent, they point out that thermal waters ascend through certain confined favourable underground channels. Manikaran, where both discharge and temperature are maximum, does not appear to overlie an extensive positive subsurface thermal anomaly. The thermal waters perhaps derive their heat from regions further away.

v) That isotope studies do not show the characteristic oxygen shift (i.e., Oxygen enrichment in thermal waters due to their association with high temperature rocks) generally associated with high temperature geothermal systems.

vi) That thermal waters of various hot springs of the Parbati valley show chemical similarities indicating that they share a common hydrothermal system, which originates somewhere near Khirganga and continues upto Jan. It appears most likely that rain and glacial melt from the mountains whose peaks ascend to over 6,000 m around the valley penetrate to depths of 4-5 km where they get heated before ascending towards the lower reaches of Parbati valley in a cluster of five springs.

vii) That the reservoir temperature determined from common chemical geothermometers (T_{SiO_2}, $T_{Na-K-Ca}$) show variations in respect of different samples collected from the valley. Processes of mixing, dilution of deep thermal waters with descending cool waters, and mineral equilibration at shallow depths appear to be quite prominent as also conductive cooling along the path of movement. Much reliance can not therefore, be placed on these estimates of the reservoir temperature in the case of Parbati valley. The reservoir temperature estimated after magnesium correction to $T_{Na-K-Ca}$ and as determined from $T_{K-Mg-Ca}$ are much lower than 200°C (Table VI-7). In the light of this and considerations discussed above the most likely reservoir temperature of Parbati valley hydrothermal systems should be around 150°C.

viii) That geological and geoelectrical investigations indicate large thicknesses of Manikaran quartzite and Pre-Cambrian gneisses. Overthrusted sheets of Kulu formation (Phyllites and Schists etc.) whilst providing favourable conditions, cannot prove to be viable reservoir rocks. The Parbati valley geothermal reservoir must therefore lie at a great depth of the order of $\geq$ 1.5 km, and is controlled by fractures in the quartzite.

ix) That in view of the deep circulation and large discharge, although of moderate temperature, appears to be a definite potential for using these thermal waters for various cultural purposes and for refrigeration and cold storage of local fruit and other produce, to which possibly electrical power generation may be added later as the development of secondary turbines becomes a reality.

6.8.8 Conclusion

The Parbati valley geothermal system owes its origin to deep percolation of snow and glacial melt as well as rain waters facilitated by tectonic fractures. The five springs of the system appear to share a common hydrothermal system, and do not show signs of any components derived from high temperature rocks.

The Manikaran reservoir is most likely located at a great depth of the order of 1500 m or more and its temperature is 150°C. A suitable location for drilling an exploratory borehole should lie around the confluence of the rivers Brahmaganga and Parbati.

CHAPTER VII

CONCLUSIONS

7.1 Geothermal Resources : Main Results and Inferences

Knowledge of probable repositories of resources, their extent and configuration is the first step towards their being exploited and developed for human good. The next step is to make some quantitative estimates of their important characteristics, if possible, without drilling. In the case of geothermal resources, these happen to be their location in space, dimension, temperature, porosity and enthalpy. The acquisition of all this information is the dream of those engaged in a geothermal exploration programme.

The study presented in this thesis demonstrates how it is possible, by analysis of available information not specifically gathered for this purpose, to make reasonable inferences about the location and probable type of geothermal resources including important features of the reservoir. We approach this problem systematically by examining all available information pertaining to the geo-tectonic evolution of different units of the Indian sub-continent and known occurrence and character of hot springs, in the light of other affine geological and geophysical information. In particular, the dramatic crustal features bounding the Indian subcontinent on the north, caused by the still continuing, albeit slower, collision of the Indian plate with the Eurasian, appear to have created conducive environments for the accumulation and concentration of geo-heat in shallow regions which constitutes potential geothermal resources of various degrees in certain parts of the Himalayan region as well as of the main land. Four main geothermal provinces where geothermal energy resources of economic viability are likely to exist have been delineated. These are:

1. The Himalayan-Burmese-Andaman-Nicobar Arc Geothermal Province.
2. The Narmada-Sone-Dauki lineament Geothermal Province.
3. The Konkan Geothermal Province.
4. The Cambay Graben Geothermal Province.

7.1.1 The Puga valley reservoir

Detailed thermal and geoelectrical methods supported by geochemical thermometry were accordingly launched to study geothermal resources of the Puga valley hot spring area which appeared to be the most promising. Upon their synthesis, results of these investigations lead to the following conclusions.

* That the valley basement is of undulating nature having its deeper parts located towards the south-west.

* That a thermally anomalous zone of about 3 sq.km area is located in the central part of the Puga valley.

* That the above hot zone is characterized by the presence of a low electrical resistivity (5 to 30 ohm.m) sub-surface formations of thickness varying from about 50 m to over 300 m.

* That a reservoir lies at shallow depths beneath the above hot zone.

* That the shallow reservoir, most likely, appears to be of a self sealing type and the flow of hot fluids probably takes place through lateral narrow channels occurring across the direction of the valley axis.

* That the temperature of the shallow reservoir is likely to be about 165°C and the porosity around 21%.

* That the shallow reservoir is a secondary one and the base temperature of the main reservoir which could not be detected even on detailed investigations, is most likely, over 200°C.

* That the Puga geothermal field is of mixed type and a small vapour/gas dominated zone, formed in the structural highs, most probably exists within the shallow reservoir.

* That the Puga geothermal field is characterised by high near surface temperatures and temperature gradients, and high heat flow (about 13 HFU).

The results of the thermal and geoelectrical surveys and investigations were supported by drill holes which tapped buoyantly upflowing thermal fluids from the shallow reservoir. A telluric survey later carried out by Rakeshkumar et al. (1979) yielded low 'J' values (relative ellipse area) within the low resistivity zone.

The thermal data of the Puga geothermal field further revealed:

i) That the natural total heat transmitted on the surface through the hot zone is around 20 billion cal./hr. which is theoretically equivalent to 20 Megawatts (MW) of electrical energy.

ii) That the probablg potential of the shallow reservoir from its energy storage along, is 2.5 MW for a 20 year period.

iii) That the Puga hydrothermal field is adequate to meet a minimum of about 5 MW electricity requirement for a period of 20 years. However, the actual life of the field at this rate or even at higher rates of exploitation should be much greater.

7.1.2 The Manikaran reservoir

The Manikaran hot springs which emerge in the Parbati valley, Kulu, Himalaya through joints and fractures in a thick sequence of massive quartzites, were the next to receive attention for a thorough investigation.

Geoelectrical surveys carried out in the area did not reveal any conductive zones upto the sensing depth of probing and persistent attempts to explore the subsurface, in the Parbati valley, using various geolectric tools, met with little success. However, it was clear from the very outset that since the resistive quartzites of the country rock were very thick and the reservoir probably existed at a great depth the most fruitful approach before exploratory drilling which would be expensive in quartzites, would be to ascertain the probable reservoir temperature by geothermometry.

The main thrust of investigations was therefore directed towards analyzing the silica and Na-K-Ca geothermometer which yielded significantly different temperatures. Tritium isotope studies of the cold and thermal waters along with their chloride contents also indicated mixing of deep upflowing thermal waters with colder meteoric waters before their emergence on the surface. A methodology was accordingly developed to ascertain the mixing ratios as well as the deep aquifer temperature, from the composition of mixed thermal waters and of cold waters of the area. This led to the result that i) thermal waters are constituted about 40% deep thermal water and 60% cold meteoric water, and that ii) the reservoir temperature is likely to be around 200°C.

Boreholes subsequently drilled in the area to medium depths, near suspected or inferred shear or contact zones, tapped thermal waters of similar temperature and composition as those obtained from the surface discharge of hot springs, thereby casting doubts regarding the temperature of the Manikaran reservoir (high around 200°C) estimated on the basis of the mixing model (this thesis section 6.8.4). A new geothermometer developed by Giggenbach (Table VI-7) and magnessium Correction to the Na-K-Ca thermometer as introduced by Fournier and Potter (1979) were therefore, also tried to make a new estimate of the probable reservoir temperature which was found to vary greatly, leading to the conclusion that:

* the base temperature of the Manikaran hydrothermal system is likely to be around 150°C and not 200°C as indicated by the mixing model and the unmodified Na-K-Ca geothermometer. Furthermore an attempt was also made to elucidate the mutual relationship if any, among the 5 hot spring zones of the Parbati valley, making use of other affine geodata. This exercise lead to the conclusion that the thermal waters of the Parbati valley appear to share a common hydrothermal system and have been diluted before being discharged on the surface, maximum dilution occurring at Kasol, followed by Manikaran and Khirganga.

* That Manikaran where the discharge and temperature

of thermal waters are both maximum does not appear to overlie an extensive positive subsurface thermal anomaly. The thermal waters seem to derive their heat from regions farther away.

* That the hydrothermal system perhaps originates somewhere underneath Khirganga area and continues up to Jan and the geothermal reservoir must be at a great depth of the order of $\geq$ 1.5 km.

7.2 Heat Flow Studies : Main Results and Conclusions

The Indian Peninsular, is generally believed to be a stable landmass. However, perturbations caused by local, regional as well as global processes continue to disturb its serenity. Effects of diversities in the initial conditions of its composite segments accentuated by the stresses produced by the continuing collisions of its northern boundary, have overprinted on its predominently ancient surface many quasi-permanent features with varying time constants. Heat flow, being directly diagnostic of thermal states of the Earth's crust and interior, a study was undertaken to analyse significant regions of the Indian subcontinent viz., the Dharwars, the Aravalli Mountain belt, the Deccan traps and the Tertiary basin of the Indian land mass. Results of these studies are briefly summarised below:

* That the Dharwars which consist of greenstone belts and some of which are equivalent to the greenstone

belts of other shields are characterised by low heat flow values 0.95 HFU.

* That the Aravalli mountain belt consisting of a sequence of Pre-Cambrian meta-sediments of three super groups, the Pre-Aravallis, Aravallis and the Delhis, once an impressive feature and now reduced to their present size by erosion, and which perhaps suffered rejuvenation in Mesozoic times, is associated with contrasting thermal structure; the Delhi Super group being characterised by high heat flow ($\approx$ 1.7 HFU); whilst the other two super groups are associated with much lower heat flow values ($\approx$ 0.95 HFU).

* That the Deccan traps which presently cover an area of about 600,000 sq.km and constitute a series of basic lava flows of Paleocene age are most likely characterised by low heat flow values typical of continental stable regions; although it must be cautioned that this is based on just two heat flow determinations at widely separated locations (vide section 3.9.3, and a new value).

* That the Cambay basin, which is an intracratonic basin, perhaps formed by discontinuous normal faults and in which a succession of ~~d~~ sediments from Recent to Eocene overlie an irregular surface of Deccan traps, exhibit a contrasting thermal structure of

its segments. The part of the basin situated towards the north of the Mahi river and in which heat flow measurements were made at six widely separated locations between Cambay and Meshana, conducts more heat (1.8 to 2.3 HFU) than its southern portion. The lowest heat flow in the basin was observed in the Broach area, where the sediments also attain a maximum thickness.

* That the association of high heat flow zone of the basin with a gravity 'high' points to the presence of a buried igneous intrusion of Pliocene to Miocene age in the basin, which probably extends from Cambay towards Kasol and Patan.

Precambrian rocks, form the basement of almost all younger formations and also contribute to most of the unperturbed component of surface heat flow. Global heat flow studies of the shields areas thus provide a useful backdrop for understanding the thermal structure of various crustal regions of the earth. An analysis of the known heat flow values in various shields and those measured over the Indian shield yields the following conclusions:

* That characteristic regions of low, intermediate and high heat flow values are found to occur in most shields.

* That the low heat flow value obtained in the high grade Archaean terrains is caused by the erosion of radioactive elements alongwith the top layers, whereas in the Greenstone belts, its existence is related to the initial conditions of their formation.

* That inspite of the widely held belief of the prevalance of steeper geothermal gradients during the Archaean and their subsequent gradual decline, there did exist segments of the crust during early Proterozoic times in which the temperature gradients were small.

* That the Proterozoic regions of the shields are characterised by higher heat flow as compared with the Archaean and the Proterosoic miogeosynclinal belts, e.g., the Delhi Super Group of the Indian Shield, have been found to mark the highest heat flow regions in the shields.

* That the lithosphere underlying such portions of the shield which experienced geosynclinal developments during the Middle Proterozoic appears to have been less ~~da~~ differentiated.

* That the Gondwanic shields in general appear to conduct higher heat flow than their Laurasian counterparts and the reduced heat flow and the mantle

component of heat flow is liekly to be different under different shields and/or their parts.

7.3 Geothermal Resources of India and the Thermal Field of the Indian Peninsula

The major economically viable geothermal resources of India happens to be of the hydrothermal type. Geopressured geothermal resources of Cambay Basin (this thesis Section 5.5.4) if properly explored could also turn out to be of economically feasible. Geothermometric studies show that about 76 hydrothermal convection systems in the Indian subcontinent are likely to possess high reservoir temperatures, constituting a total stored heat potential of about 25×10^{18} cal. 34 of these systems are likely to possess reservoir temperatures exceeding 150°C and can therefore be utilized for power generation with a total estimated power potential of 385 MW centuries or 1280 MW for a 30 year period of utilisation. The remaining 42 systems are of intermediate temperatures, ranging from 80 to 150°C and could be utilized for other cultural and industrial purposes, pending the development of binary turbines which should eventually render them as viable sources of electrical power (Gupta and Drolia, 1977; Hari Narain and Gupta 1978). Most of the high temperature reservoirs (temperature 150°C) are found in the Himalayan region, but those in Agnigundala (Godavari Graben), Tatapani (Surguja Dist, M.P) and Surajkund (Huzaribagh dist. Bihar) also appear to hold good prospects for encountering storing high enthalpy fluids suitable for power generation.

Beside the four tectonic units in which heat flow measurements were made (Chapters III and IV A) some investigations have also been made in the following type areas (Fig VII-1) of the subcontinent:

1. Peninsular Gneisses
2. The Singhbhum thrust zone
3. The Bijawars (Precambrian)
4. The Cuddapah basin
5. Various Gondwana sedimentary basins
6. The Assam basin
7. Some locations in the Himalayan Orogen

Heat flow values over the Indian Peninsula, excluding areas overlying hydrothermal conventive systems, show considerable variation from 0.6 to 2.5 HFU. The last known thermal/magmatic event dates back to Late Cretaceous or Lower Eocene. During a tectono-thermal event, parts of the crust and upper mantle gets thermally perturbed; by the processes resulting in magmatism, metamorphism, deformation and uplift. The time taken for the perturbation is its contribution to surface heat flow to die out depends upon the depth upto which the upper mantle gets disturbed. According to Vitorello and Pollack (1980) such perturlations in Cenozoic tectonic zones account for approximately 30% of the heat flow, whereas they are negligible in late Precambrian to mid-precambrain terrains.

The outpouring of the Deccan lavas is the most promising Cenozoic thermal event which happened to the

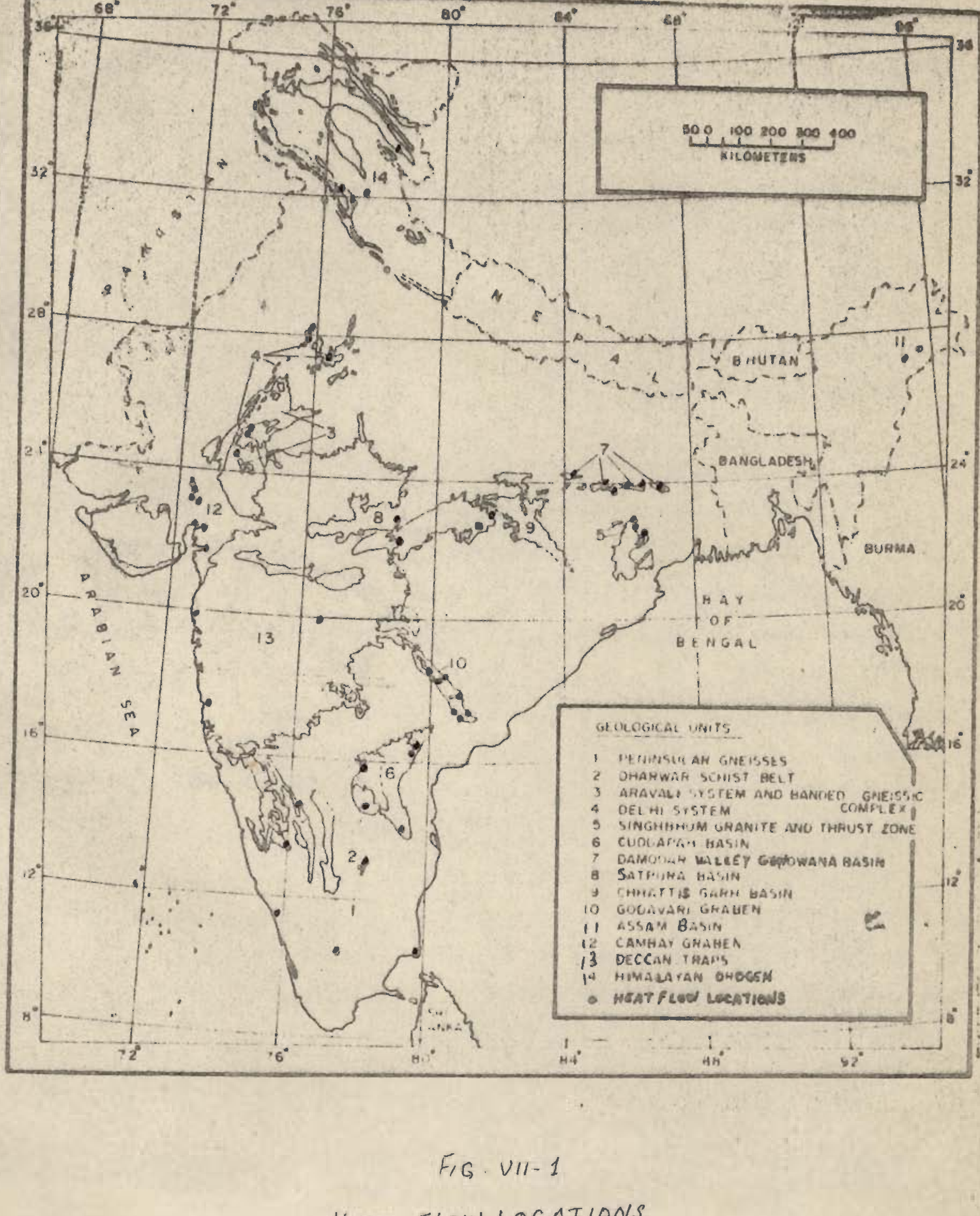


FIG. VII-1

HEAT FLOW LOCATIONS

Indian Peninsula. However two heat flow values determined over the Deccan Traps yielded normal values typical for stable continental platforms. The only plausible explanation appears to be that the Indian Shield riding a thick lithosphere, after its detachment from the Gondwanaland traversed a hot spot that gave rise to Deccan Trap volcanism with the effusive centres shifting from North to South. The volcanic material spread over the surface subsequently cooled losing all its original heat of the lama. However over and around the feeder channels, which must still be relatively ward, the temperature gradient should be higher.

The Bijawars of Central India and the Singhbhum thrust zone are associated with low to intermediate heat flow values. The variation over these areas as well as in Peninsular Gnessic terrains, most part of the Cuddapah basin and the Araullis can be attributed mainly to, a) the differing initial conditions obtaning at the time of their formation, such as the degree of partial differentiation, and migration of the heat producing elements K, U, Th from the deeper levels and their redistribution in the lithosphere, and b) vertical movements, and erosion studies just completed before writing these lines have indicated that the mantle under lying the Dharvar Craton (Chepter IV C) as well as the central part of the Indian Shield conducts more or less the same amount of heat flow. The main variation in the surface heat flow is caused by the variation in heat production in a top thin layer of the crust. The situation is not clear for the northeastern margin of the Cuddapah basin where a high heat flow value (1.8 HFU)(Chapter IV B) has been reported. Rao et al (1976) suggest a heat flow provines for the Proterozoic Shield of India, with a high value for the reduced heat flow. It seems possible that a very wide scatter occurs in the distribution of radioactive heat generating elements in the upper crustal layers of various regions of Peninsular India. It will give rise to different values for the intercept and slope of the straight line representing heat flow and heat generation. Therefore much more extensive heat flow and heat generation studies are called for.

Gondwana basins (Damodar valley and SE Part of the Godavari valley) exhibit high heat flow values. The causes of high heat flow in these rift valleys appear to be, a) enrichment of basement rocks in radioactive isotopes; b) water circulation in sediments and c) transient thermal perturbation introduced due to the probable subsequent rejuvenation. However, a representative picture, inspite of a large data set, is yet to emerge for these areas.

In the case of Peninsular India, it is our opinion that the heat flow variations are confined to the lithosphere and that the quantity of head flowing upward from the asthenosphere is uniform. Consequently lateral variations in subsurface temperatures within the lithosphere do exist. These cause thermal stresses which are sometime released through seismic activity along ancient zones of weakness.

7.4 Suggestions

The Indian land mass in many ways represents a unique segment of the lithosphere. Apart from the long translational journey northwards since the breakup of the Gondwanaland, it has suffered significant rotation, rollision and underthrusting, till tale marks of which are widely, although not always illuminatingly, manifested. Thus with the evolution of the Himalaya and the vast-Indo-Ganga-Brahmputra sedimentary basin, which was created in its wake, vertical movements also occured in various parts of the Indian Peninsula. The attendent large scale transport of matter and consequent redistribution of the thermal energy regime in the subcrustal layers should no doubt be reflected in the heat flow pattern over the Indian subcontinent. A systematic study of the thermal field in as many as possible geotectonic units of the Country interpreted in the light of other geo-information is most likely to provide results of great geodynamic significance regarding the evolution of not only the Indian Shield but, the continental crust generally. For example it is believed that a part of the Aravalli Precambrian crust shank and was covered with later deposits. How are the precambrian and the later sigments of the Aravalli craton related and flow values should suggest useful clues to these.

Heat flow investigations could be especially valuable for elucidating the tectonically active region of the Northeastern India where high stress accumulations that manifest themselves so eloquently through abnormal seismic activity, is expected to be characterised by equally dramatic thermal regimes.

Geothermal Exploration

Equally desirable are reconnaissance geological and hydrogeological studies of hot spring areas supported by detailed geochemical investigations of the thermal and cold waters and isotopic studies. Analysis also needs to be made to delineate parameters intrinsically related to reservoir temperatures of a hydrothermal system, evaluated based on experimentally worked out models of hot water rock interaction. Once this is done, extensive geophysical surveys should follow to pin-point the location, delineate the shape and size of a geothermal reservoir. Recource should be made, if necessary to laboratory studies of electrical and other physical properties of the region under simulated field conditions (water saturation and temperature). Such a system approach will lead to a proper understanding of the field data and conceptualization of more realistic models of a geothermal reservoir. Detailed grological mapping aided by slim drill holes should then enable one to determine reservoir characteristics and its potential precisely enough for an economic undertaking.

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