SURFACE WATER AND GROUNDWATER INTERACTION STUDIES USING ISOTOPE TECHNIQUES

A THESIS

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CANDIDATE'S DECLARATION

I hereby certify that the work, which is being presented in the thesis entitled **"Surface water and groundwater interaction studies using isotope techniques"** in fulfilment of the requirement for the award of the degree of Doctor of Philosophy, submitted in the Department of Earth Sciences of this University, is an authentic record of my own work carried out during the period from **August, 1993** to **July, 2000** under the supervision of **Dr.Rm.Manickavasagam, Dr. Bhishm Kumar** and **Dr. S. V. Navada**.

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

Dated: 28.07.2000.

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

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ABSTRACT

Fresh water available from surface water bodies, such as, lakes and rivers, is a precious resource for mankind. As fresh water is becoming increasingly soarce in recent times, it needs proper attention. Improper management of fresh water has resulted in extensive environmental degradation of many lakes and rivers in India. The environmental damages caused to surface water bodies, may propagated further into hydraulically connected groundwater system and would drastically reduce the availability of this alternative source of potable water.

Surface water and groundwater interaction studies provide useful information about environmental impact and for better management of available fresh water resources. In this present study, an attempt has been made to study surface water and groundwater interaction by using the stable isotopic technique in two different hydrogeological settings, namely, the Nainital lake (mountain terrain) and the upper reaches of River Ganga (alluvial riverine system) in western Uttar Pradesh. These water bodies have been selected for the present investigation, as they form the main fresh water resource for large area of irrigation besides domestic use.

Nainital Lake - Groundwater Interaction

Nainital lake (29° 23' 09" N and 79° 27' 35" E) is a high altitude (1937 m above m.s.l.) natural lake located in Nainital district of Uttar Pradesh. It is a crescent - shaped lake with maximum length and width of 1.4 km and 0.45 km respectively and maximum and mean depths of 27.3 m and 18.5 m respectively. The surface area of the lake is 0.46 Km^2 with a maximum capacity of 8.57 Mm^3. The average annual rainfall in the basin is 2488 mm and is geologically characterized by Krol and Tal formations, which are folded and faulted.

Direct precipitation over the lake has been estimated by Theissen Polygon method using rainfall data collected from four raingauges installed in the lake catchment. The surface inflow to the lake, due to rainfall has been estimated by two different methods, viz. Soil Conservation Service - Curve Number method (SCS-CN) and Lake Level Trend Analysis method (LLTA). The latter method was developed in the present study. SPOT satellite imagery, ERDAS™ software, Survey of India toposheet (#53 O/7) and Nainital guide map were used for obtaining land use

information, which in turn was used in the selection of appropriate runoff curve number for SCS-CN method. The weekly rates of reduction of water level in the lake level were considered for the LLTA method. Surface inflows into the lake through the drains were measured at discrete time intervals, with an assumption that the flow in the drains does not vary significantly during intermittent period. The change in storage of the lake was estimated from daily lake level records. Surface outflow from the lake was computed by empirical method applicable for submerged rectangular sluice openings. Evaporation from the lake was computed from the meteorological data collected near the lake by modified Penman method.

Water samples collected during 1994 - 1996 were used for chemical and isotopic characterization of the springs and the lake. A total of 63 samples were collected from the springs, 273 from Nainital lake at different locations and depths, and 39 from the drains for chemical characterisation. Major ions were analysed by standard procedures. The $\delta^{18}O$ and δD isotopic characterisations were done for local rainfall (16 samples), drains (7 samples), Nainital lake (184 samples) and springs (35 samples). Isotopic composition of the lake evaporates was calculated by Craig and Gordon model. $\delta^{18}O$ was used as tracer to calculate the proportion of the lake water drawn by the wells. Sub-surface components of the lake were estimated by isotope mass balance method, by using isotopic data in conjunction with conventional data. Estimates of uncertainties in the measured / estimated values of a few water balance components were made based on literature. The propagated errors in the estimated sub-surface components were evaluated by standard methods.

It was noted that the lake temperatures during December to February vary from 8° to 10°C both in epilimnion and hypolimnion zones, indicating that the lake is thermally well-mixed and homogeneous and the thermocline, which had developed in March disappears totally in November. During the period from March to November, the temperatures in epilimnion zone vary from 15° to 25°C, whereas in the hypolimnion zone, they vary from 8° to 10°C, identical to the range that was observed in winters.

In winters, $\delta^{18}O$ in epilimnion and hypolimnion zones, ranges from -8.2‰ to -9.7‰ and -8.1‰ to -9.8‰ respectively, whereas δD remains -49‰ and ranges from -52‰ to -55‰ respectively, indicating that the lake is isotopically homogeneous and well-mixed. During summers, $\delta^{18}O$ in epilimnion and hypolimnion zones varies from -5.5‰ to -8.1‰ and -6.6‰ to -7.7‰ respectively, and the δD varies from -38‰ to -53‰ and -46‰ to -50‰ respectively, suggesting that during summer, the effect of evaporation is pronounced in the epilimnion zone and during monsoon stratification renders the warmer surface inflow, float over the cooler hypolimnion zone.

Isotopic investigations carried out in Nainital lake catchment reveal that the weighted annual mean $\delta^{18}O$ of rainfall is about -11.3‰. The estimated altitude effect on the rainfall in Nainital area in $\delta^{18}O$ is about -0.34‰ and in δD about -2.4‰.

A multiple linear regression model has been developed for generating $\delta^{18}O$ data of rainfall in Nainital area, by employing the meteorological parameters such as, monthly mean relative humidity, monthly rainfall and monthly mean air temperature.

δ_E was estimated by Craig and Gordon Linear Resistance model. The results of the model indicate that δ_E is controlled mainly by the relative humidity. The estimated δ_E is heavier during months of higher relative humidity, as compared to the months of lower relative humidity. A lower limit for isotopic enrichment of a fresh water lake has been identified by analysing the Craig and Gordon Linear Resistance model. The lower limit is determined by the difference between isotopic composition of the lake (δ_L) and that of the atmospheric vapour above it (δ_a). The evaporative enrichment of the water body will continue if (δ_L - δ_a) is greater than the equilibrium enrichment factor.

The total cations of the springs, normalised to those of the lake, suggest that the springs located in Balia ravine (downstream side of the lake), are hydraulically connected to the lake. This is confirmed by isotopic data obtained for Balia ravine, where during summers and monsoons $\delta^{18}O$ and δD values remain -5.9‰ to -8.8‰ and -55‰ to -64‰ respectively and during winters, they remain -7.9‰ to -9.5‰ and -44‰ to -56‰ respectively. Data pertaining to summers and monsoons, indicate that the epilimnion zone of the lake is the main contributing source for Balia ravine springs. It implies that reduction in the lake level below the present level, might result in drying up of the springs in Balia ravine.

Conventional water balance studies, carried out for estimating inflow components show that the groundwater contributes 50%, surface runoff from the catchment 30%, drains 10% and direct rainfall over the lake surface contributes 10% of the total annual inflow. Similarly, studies carried out for estimating the outflow components indicate that sub-surface outflow accounts for about 55%, evaporation 10% and surface outflow for about 35% of the total annual outflow. Annual pumping data indicate that outflow through seepage towards northern bank of the lake is about 40% of the total annual outflow from the lake. This is substantiated by the long-term surface outflow and annual rainfall data analyses, which indicate reduction in the surface outflow for a given amount of annual rainfall in the past three decades. It is, therefore, inferred that any change in the quantity of pumping will affect the availability of water in the lake.

Hydrogeological investigations indicate that the shale formation, which occupies about 50% of the lake catchment area, has an hydraulic conductivity of about $5.4 * 10^{-8}$ m/s and a specific yield of about 0.015%, suggesting it is not a suitable aquifer. However, the catchment area has well developed lineaments and faults. The hydrologic investigations conducted along these lineaments indicate higher infiltration capacity (about 58 cm/h). Therefore, it is inferred that most of the groundwater inflow to the lake might be occurring along these zones. Further, due to higher infiltration capacity of Sukhatal lake, its seepage appears to be a major recharge source for Nainital lake and therefore, any activity in Sukhatal lake catchment may affect the water quality and availability in Nainital lake.

The slope of the $\delta^{18}O$ - δD water line of the lake is 7.1, which is very close to the Local Meteoric Water Line of 7.5, indicating that the lake is rainfall dependent and any change in annual rainfall might be reflected on the isotopic characteristics of the lake.

Water retention time was computed for the Nainital lake by different methods. The isotopic mass balance method gives 1.93 years, chloride mass balance technique 1.77 years and the conventional water balance method yields 1.92 years. Sensitivity analysis carried out for the isotope mass balance method, indicates that the method is highly sensitive to the difference between the $\delta^{18}O$ values of groundwater inflow and that of lake seepage. It was also found that the relative error decreases with increase in the difference between these two isotope indices.

River Ganga - Groundwater Interaction

The River Ganga flows in one of the largest alluvial basins of the world and the enters the plains at Hardwar after flowing nearly 260 km in hilly terrain. For the present study, the river reach between Hardwar and Narora of western Uttar Pradesh has been chosen. The study area falls between 28° to 30° N latitudes 77°45' to 78°15' E longitudes. This reach of the river, caters water to extensive land developed for agricultural use through four major canals, namely, Upper Ganga Canal, Madhya Ganga Canal, Lower Ganga Canal and the Eastern Ganga Canal.

Groundwater levels were observed using the piezometers installed at Balawali, Rawalighat, Brijghat, Anupshahr and Rajghat (near Narora). Groundwater hydraulic gradients were estimated from this data. Single well tracer dilution experiments were conducted to estimate the hydraulic conductivity. Specific discharge of groundwater to the river at different sites were estimated by Dupuit's method. Two- and three- component mixing models were used to estimate groundwater discharge to the river. The rainfall isotopic index for the study area has been computed by using long-term (1961-1995) monsoon isotopic data of New Delhi station.

For isotopic characterisation, 63 water samples from river Ganga and 122 groundwater samples from handpumps existing closer to the river were collected during November, 1992 to May, 1995. Another 35 groundwater samples were also collected from either side of the river at a distance of 1, 5 and 10 km perpendicular to the river course, during January 1994 and analysed for their isotopic characteristics..

It has been found from the groundwater level measurements that both western and eastern aquifers contribute to the river flow at Balawali, Brijghat and Anupshahr, in the non-monsoon seasons. Whereas, $\delta^{18}O$ data of groundwater in the eastern aquifer near Anupshahr, shows that the influence of the river decreases with increase in distance. At Rajghat (Narora), it appears from the groundwater level data that the river receives groundwater from the western aquifer, and seeps towards the eastern aquifer during the same period. These indicate that the nature of river and groundwater interaction is different in different reaches of the river, in the study area.

The $\delta^{18}O$ variation in the river at Brijghat, Anupshahr and Rajghat, during pre-monsoon seasons is comparable to the variation observed in western aquifer, and therefore, the western aquifer seems to be a major source of water to the river during pre-monsoon period.

Analysis of the rainfall isotopic data from New Delhi station shows that the long-term monsoon weighted average for $\delta^{18}O$ is -6.4‰. Therefore, it is likely that the groundwater $\delta^{18}O$ in the study area might be around -6.4‰.

The frequency distribution analysis of $\delta^{18}O$ were carried out by using samples pertaining to groundwater collected closer to the river, during pre-monsoon season. The results show the western and eastern aquifers have different peak values viz. -8.5‰ and -9.5‰ respectively. It indicates that different processes are involved in influencing groundwater isotopic characteristics of the two aquifers. Spatial variation in $\delta^{18}O$ values of the aquifers is considerable and it is due to various hydrological processes. The western aquifer is recharged mainly by precipitation, seepage from Upper Ganga Canal, and also by the irrigation return flow. On the other hand, the eastern aquifer is recharged mainly by precipitation, seepage from Eastern Ganga Canal (restricted only to the northern part of the study area), Ram Ganga Feeder Canal and in certain zones by the river Ganga.

The frequency distribution analysis of $\delta^{18}O$ were carried out by using samples pertaining to groundwater collected closer to the river, during monsoon and post-seasons. The results show the western and eastern aquifers have different peak $\delta^{18}O$ values of -6.5‰ and -5.5‰ respectively. The enrichment of $\delta^{18}O$ during these seasons reflects the influence of rainfall recharge to the groundwater. The enrichment is also possible due to evaporation of groundwater, as groundwater table closer to river at many sites which are found to be less than one to two metre below the ground level.

A plot of $\delta^{18}O$ and δD pertaining to river Ganga and those pertaining to New Delhi rainfall shows that there is no discernible enrichment effect on isotopic characteristic of the river. This signifies that evaporation may not be a major factor for isotopic variations in the river reach between Hardwar and Narora.

The mean $\delta^{18}O$ values indicate a progressive enrichment in $\delta^{18}O$ during pre-monsoon season from Hardwar to Brijghat river Ganga, whereas it becomes more negative between Brijghat and Rajghat. It is observed from the isotopic data of both river and aquifer systems that the enrichment of $\delta^{18}O$ in the river could be due to the contribution of groundwater from western aquifer rather than from eastern aquifer.

The specific discharge of groundwater into Ganga river between Hardwar and Balawali has been computed using two component mixing model and it ranges from 13.4 m^2/day in June 1993 to 26.8 m^2/day in March 1994. These are much higher than those calculated by Dupuit's method. This discrepancy may be due to linear alignment of the piezometers normal to the flow direction of the river and consequent estimation of the apparent hydraulic gradient of groundwater than true gradient.

Groundwater contribution to river Ganga between Rawalighat and Brijghat has been estimated by using a three-component mixing model. The results compare well with that of channel water balance method. This substantiates the conjecture that the $\delta^{18}O$ value of groundwater occurring close to the river, particularly when the groundwater level is higher than the river level, is the appropriate isotopic index of groundwater.

The contribution of groundwater to the flow in river Ganga at Balawali, Anupshahr and Rajghat during non-monsoon season ranges from 21% to 67% at different sites, while during monsoon season it ranges from 17% to 60%. The quantity of groundwater in the river at Brijghat varies from 5% to 76% during non-monsoon season, and from 18% to 35% during monsoon season. During monsoon season, seepage from river Ganga is retained as bank storage, which subsequently flows back into the river.

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At each and every step of my life, I feel The Protecting Hands of The Almighty. I bow at His Lotus Feet for the grace and blessings being showered on me.

Rm. P. Nachiappan

LIST OF PUBLICATIONS FROM THE PRESENT WORK

1. Bhishm Kumar and Rm P Nachiappan, 1999. **On the sensitivity of Craig and Gordon model for the estimation of the isotopic composition of lake evaporates.** *Water Resour. Res.* 35(5):1689-1691.

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CHAPTER 1

INTRODUCTION

1.0 INTRODUCTION

Fresh water available from the surface water bodies such as lakes and rivers is a precious resource for the mankind that needs proper attention, as it is becoming increasingly scarce in recent times. Improper management of fresh water has resulted in extensive environmental degradation of many lakes and rivers in India. The environmental damages caused to surface water bodies may further be propagated into the hydraulically connected groundwater system that would drastically reduce the availability of this alternative source of potable water.

In case of riverine systems, groundwater has been recognized as not only the source for base flow in the streams but also an important contributor during storm flow (52, 136). In recent years, the hydrologists have been focusing on the physical mechanism of runoff generation (48, 79). However, it is the quantum of interaction between surface water and groundwater system that is crucial for the water resources managers, as the water availability in rivers and lakes has direct relevance on irrigation and domestic water supply projects.

The surface water and the groundwater systems have fairly been well understood independently but interaction between them continues to be an active area of research. Improvements in understanding the interaction between surface water and groundwater may facilitate in the better management of available fresh water resources (161). Although, analytical and numerical models of surface water and groundwater interaction are being increasingly used, they need continuous improvements for effective management of water resources in different environmental set-up. The future studies will also be benefitted from increased use of interdisciplinary approaches such as isotope geochemistry (159). Although, the isotope techniques have been applied to surface water and groundwater interaction studies, they are yet to be comprehensively tested under diverse environmental and hydrogeological conditions.

The isotopic technique has been applied to several lakes worldwide to understand the lake-groundwater interaction (39, 51, 82, 91). However, it has been applied mostly to low or mid altitude lakes that are hydrologically and isotopically in a steady state condition. In India, the lake-groundwater interaction study has not been attempted so far to any of the lakes located either in the mountainous region or in the plains by using stable isotope technique.

In view of the above, an attempt has been made to study the surface water and groundwater interaction by using the stable isotopic technique in different hydrogeological settings namely the Nainital lake (mountain terrain) and the upper reaches of river Ganga (alluvial riverine system) in western Uttar Pradesh. These water bodies have been selected for the present investigation as they form the main fresh water resource for large area of irrigation besides domestic use.

Lake Nainital (29° 23' 09" N and 79° 27' 35" E) is a high altitude (1937 m above m.s.l.) natural lake located in Nainital district of Western Uttar Pradesh. It is a bean shaped lake with a maximum length and width of 1.4 km and 0.45 km respectively, and having a maximum and mean depths of 27.3 m and 18.5 m respectively. The surface area of lake is 0.46 km^2 with a maximum capacity (volume) of 8.57 Mm^3. The average annual rainfall in the Nainital basin is 2488 mm and is geologically characterized by Krol and Tal formations that are folded and faulted.

River Ganga flows in one of the largest alluvial basins of the world and enters the plains at Hardwar after flowing nearly 260 km in the hilly terrain. The 220 km long river reach between Hardwar and Narora caters water to four major canals namely Upper Ganga Canal (UGC), Madhya Ganga Canal (MGC), Lower Ganga Canal (LGC) and the Eastern Ganga Canal (EGC) which have been developed mainly for agricultural use. In view of its importance, the river reach between Hardwar and Narora (28° to 30° N latitudes and 77°45' to 78°15' E longitudes) in western Uttar Pradesh has been chosen for the present study.

The river-groundwater interaction study has earlier been conducted in some parts of the river Ganga. The isotopic investigation on part of the Gangetic basin lying in Bangladesh revealed that the river has no influence on groundwater, but based on hydrometric data it was observed that the base level of groundwater and the water table is determined by the river (43). The stable isotope ($\delta^{18}O$ and δD) characteristics of the river from its origin to Patna (around 1300 km downstream of Narora) indicate that the river is progressively enriched in its isotopic composition along its course due to inflow of water from tributaries and also by groundwater besides evaporation loss (120).

The regeneration of river Ganga in the study area has earlier been investigated a) by using river discharge data (145), b) by stable isotope technique (105) and c) by comparing the river and groundwater levels (146). It has been reported that the river is effluent during non-monsoon season (145,146). The findings of the earlier isotopic study (105) were not compared with river discharge data but the investigators reported that the contribution of groundwater to river Ganga during non-monsoon months resulted in the progressive enrichment of river isotopic composition in the downstream direction.

In the present investigation, an attempt has been made to employ isotopic technique in conjunction with conventional methods to understand the nature of interaction in the Nainital lake with groundwater system and river Ganga with groundwater system.

1.1 Objectives

The main objectives of the surface water and groundwater interaction study using $\delta^{18}O$ and δD isotopes for the two proposed areas are as follows:

Lake Nainital

i) To identify and quantify the major recharge sources for the lake

ii) To identify and quantify the interconnection between the lake and downstream springs

iii) To compute the water balance of the lake

River Ganga between Hardwar and Narora

i) To quantify the groundwater discharge to the river during non-monsoon season

ii) To investigate the effluent / influent nature of river Ganga during monsoon season

1.2 Methodology

Isotopic and chemical tracer mass balance methods were used in addition to Water balance method to study the Nainital lake and groundwater interaction. Estimates of uncertainties in the measured/estimated values of a few water balance components were made based on literature (158). The propagated errors in the estimated sub-surface components have been evaluated by standard methods (15).

To study the river Ganga and groundwater interaction, river flow separation has been carried out by using two- and three- component mixing models. In addition, channel water balance and Dupuit's method (49) were also adopted to estimate the proportion of groundwater in the river discharge.

1.3 Data

Two different types of data have been collected for the present investigations, which include primary and secondary data. Primary data has been obtained by using different tools and techniques. The primary data was acquired through several hydrological investigations and measurements, made in the study area during the study period (1994-96), and pertains to the various components of the water balance of the lake including chemical, isotopic and hydrometeorological characteristics of the lake basin. The primary data collected in the river Ganga and groundwater interaction study during the study period (1992-95) include the groundwater level, isotopic characteristics of river Ganga and those of groundwater. Secondary data pertain to historical records of lake and river levels, bathymetric data of lake, pump and sluice operation data, discharge data of springs and the river Ganga. These secondary data have been obtained from published and unpublished literature, Government records etc.

1.3.1 Hydrological Data

In the Nainital lake - groundwater interaction study, direct precipitation over the lake has been estimated by Theissen Polygon method for the rainfall data of four raingauge stations in the lake basin. The surface inflow to the lake, due to rainfall has been estimated by two different methods viz. Soil Conservation Service - Curve Number method (SCS-CN) and Lake Level Trend Analysis method (LLTA). The latter method, was specifically developed for the Nainital lake basin, in the present study. SPOT satellite imagery, ERDAS software, Survey of India toposheet (#53 O/7), and Survey of India guide map of Nainital area (scale 1:10,000) were used for obtaining landuse information that are required in the selection of appropriate runoff curve number for SCS-CN method. The mean weekly rates of fall in the lake level were considered for the LLTA method. The surface inflow to the lake through the drains was measured by using pygmy current meters at discrete time intervals, with an assumption that the flows in the drains do not vary significantly during intermittent period.

The change in storage of the lake was estimated from daily records of lake level. The lake level records were obtained from Uttar Pradesh Public Works Department (UPPWD). Surface outflow from the lake was computed by an empirical method applicable for submerged rectangular sluice openings, using the sluice operation records supplied by UPPWD. Evaporation from the lake was computed from the meteorological data collected near the lake by using modified Penman method (75). The groundwater draft by the wells located in the periphery of the lake has been calculated using the pump operation data supplied by Uttar Pradesh Jal Nigam, Jal Sansthan and Indian Army authorities. The proportion of the lake water drawn by the wells have been estimated using tracer techniques. Discharge of some of the downstream springs were monitored using standard measuring jars and stop watch. Historical record of discharge data of springs located in the downstream direction of the lake were obtained from UPPWD Authorities.

In the river Ganga - groundwater interaction study, groundwater level observations were made in the piezometers installed at Balawali, Rawalighat, Brijghat, Anupshahr and Rajghat (near Narora). Hydraulic gradients were estimated from the observed groundwater level data. Single well tracer dilution experiments were conducted to estimate the hydraulic conductivity. The specific discharge values at different sites were estimated by using Dupuit's method. The river discharge data at Hardwar, Rawalighat and Brijghat and the river level data at Balawali, Rawalighat and Brijghat were obtained from UP State Irrigation Department and Indian Railway Authorities.

1.3.2 Hydrochemical Data

In the Nainital lake - groundwater interaction study, chemical characterizations of the springs occurring in the study area and the lake have been carried out. A total of 63 samples were collected from the springs, 273 samples from Nainital lake at different locations and depths, and 39 samples from the drains. The physico-chemical properties such as pH and temperature were measured in situ at the time of sample collection using standard probes.

As no hydrochemical analyses were conducted for the river - groundwater system, sampling was not carried out for chemical analyses.

1.3.2.1 Sampling procedures

The water samples for chemical analyses were collected from the lake by using a standard depth water sampler of 2 litre capacity. Each water sample was filtered by using a 0.45μ filter paper before being stored in three different containers out of which one was preserved with HNO_3 for Na^+ and K^+ determination, the second was preserved with H_2SO_4 for total and calcium hardness determination and the third was unpreserved for sulphate, alkalinity and chloride analyses.

Groundwater samples for both isotopic and chemical analyses were collected from springs and pumping - wells by directly tapping the discharges as closer as possible to their issue points to avoid contamination effects. Samples from drains were collected near the drain axis and sufficiently below the surface. The samples collected from the springs, wells and drains were preserved and stored in a similar fashion as lake samples.

1.3.2.2 Chemical analyses

The chemical analyses were carried out using standard procedures (4). The chemical characterization has been carried out only for major cations such as Ca^{2+}, Mg^{2+}, Na^+ and K^+ and anions HCO_3^-, SO_4^{2-} and Cl^-. Chloride was estimated by Argentometric method in the form of silver chloride. Alkalinity was determined by Titrimetric method using phenolphthalein and methyl orange indicators, within 24 hours from the time of collection. Total hardness and calcium hardness were determined by EDTA Titrimetric method while magnesium hardness was calculated by deducting calcium hardness from total hardness. Calcium ion concentration was calculated by multiplying calcium hardness (in mg/L) with 0.401 while magnesium ion concentration by multiplying magnesium hardness (in mg/L) with 0.243. Sodium and potassium were determined by flame-emission method using Flame Photometer. Nitrogen in the form of nitrate was determined in a UV/VIS Spectrometer. Phosphate was estimated by Stannous chloride method in the form of molybdenum blue in a UV/VIS Spectrometer. Sulphate was determined by Turbidimetric method in the form of barium sulphate crystals using a standard

Turbdity meter. The overall accuracy of the chemical analyses as determined by replicate measurements and electro-neutrality, was found to be better than ±10%.

1.3.3 Isotopic Data

In the Nainital lake - groundwater interaction study, the $\delta^{18}O$ and δD isotopic characterization have been carried out for 16 rainfall samples, 7 drain water samples, 184 lake samples and 35 spring samples. The isotopic composition of the lake evaporates were calculated by employing the linear resistance model (28).

In the river Ganga - groundwater interaction study, for isotopic characterisation 63 samples from river Ganga and 122 groundwater samples from the existing hand-pumps located closer to the river were collected. In addition to the above, another 35 groundwater samples were collected from either side of the river at a distance of 1, 5 and 10 km, during January 1994. The rainfall isotopic index for the study area was computed by using the long-term (1961-1995) monsoon isotopic data of New Delhi station, obtained from the IAEA/WMO Global Network for Isotopes in Precipitation (GNIP) database.

1.3.3.1 Sampling procedures

Samples for isotopic analyses were collected from the lake, drains and springs using the same procedures as for hydrochemistry but without any preservation. In addition, rain water samples were collected from four stations set up at different altitudes, using plastic containers and metal funnels with an effective catch-diameter of 210 mm. The daily precipitation collected during each calendar month was thoroughly mixed to yield a monthly-integrated sample. All the samples for isotopic analyses were stored in 60mL High Density Poly Ethylene (HDPE) vials and sealed tightly to avoid evaporation.

For the river Ganga - groundwater interaction study, samples for isotopic analyses were collected from the river Ganga by using a grab sampler. The samples were collected sufficiently away from the bank and well below the river surface to avoid contamination effects. Groundwater samples from unconfined aquifers were collected by using the hand-pumps that normally extend to a depth of 10 to 20 m below ground level. Before collection of samples, the hand-pumps were operated for sufficiently long duration and pumped approximately three times the volume of the well, to obtain reasonably representative samples. The samples were stored as described earlier.

1.3.3.2 Isotopic analyses

Most of the samples collected for stable isotope measurements were analysed at Isotope Hydrology Laboratory, Bhabha Atomic Research Centre, Mumbai, India. Few samples of the lake were analysed at Environmental Isotope Laboratory, University of Waterloo, Ontario, Canada and a few samples of the river were analysed at the Isotope Laboratory, Centre for Isotope Studies, University of Paris-Sud, Paris, France. The samples were analysed for $\delta^{18}O$ by the CO_2 equilibration method (47) and for δD by passing through zinc shots (24). The overall analytical accuracy for $\delta^{18}O$ was found to be within ±0.2‰ and for δD within ±1.0‰.

1.4 Organisation of the thesis

The studies carried out to meet the stipulated objectives have been presented in five chapters. This chapter one on introduction includes the objectives of the study, methodology and an account of procedures adopted for the collection of hydrological, hydrochemical and isotopic data relevant to the study.

Chapter 2.0 gives an account on the applicability of isotope technique for studying the surface water - groundwater interaction, along with a brief description of the basic theory, nomenclature and different approaches to interpret the isotope data in similar studies.

Chapter 3.0 deals with the interaction of Nainital lake with groundwater system. A conceptual model for the lake - groundwater interaction has been proposed. The results of the detailed hydrological, hydrochemical and isotopic investigations are presented and the characteristics of different components of the water balance are discussed. Finally, the results of the water balance method, chemical and isotopic mass balance methods are compared and discussed.

Chapter 4.0 deals with the river Ganga and groundwater interaction study. Isotopic characteristics of the river and the groundwater are discussed in hydrological context. The results of the isotope mass balance, channel water balance and Dupuit's method have been presented and discussed.

Chapter 5.0 includes the conclusions drawn from the lake - groundwater and river - groundwater interaction studies. Further a small note, on the scope for furture studies in the application and development of stable isotope techniques, has been presented.

CHAPTER 2

USE OF STABLE ISOTOPES IN SURFACE WATER AND GROUNDWATER INTERACTION STUDIES - A REVIEW

2.0 USE OF STABLE ISOTOPES IN SURFACE WATER AND GROUNDWATER INTERACTION STUDIES - A REVIEW

Environmental isotopes are being used increasingly as tools to study various processes in hydrology. Environmental isotopes include both stable and radioactive isotopes, which occur in the environment in varying concentrations. Among the stable isotopes, oxygen and hydrogen have wide application in hydrology because, as part of the water molecule itself, they act as natural tracers. Comprehensive reviews on the applicability of the stable isotopes in river studies (55) and lake studies (63) give a basic insight to the potential use of isotope techniques.

2.1 Nomenclature

Oxygen has six isotopes viz., ^{14}O, ^{15}O, ^{16}O, ^{17}O, ^{18}O and ^{19}O. Since ^{14}O, ^{15}O and ^{19}O are radioactive with very short half-lives, they are not useful for any meaningful hydrological study. Among the remaining three isotopes, the mass abundance of ^{16}O (99.759%) and ^{18}O (0.204%) are higher than ^{17}O (0.037%). Therefore, the ratio of $^{18}O/^{16}O$ is important for hydrological studies (59). Hydrogen has three isotopes viz., ^{1}H (Protium) , ^{2}H or D (Deuterium) and ^{3}H (Tritium). ^{1}H is the most abundant hydrogen isotope (99.9855%) followed by ^{2}H (0.0145%). Tritium is a radioactive isotope with a half-life of 12.43 years. In isotope hydrology, only $H_2{}^{16}O$, $H_2{}^{18}O$ and $HD^{16}O$ are generally considered amongst various species of water molecules .

2.2 Delta (δ) Notation

The stable isotopic ratios are expressed as deviation with respect to a standard such as V-SMOW (Vienna - Standard Mean Oceanic Water), denoted by δ and expressed in permil (‰).

$$\delta_x \text{‰} = \frac{R_{x\,(Sample)} - R_{x\,(Standard)}}{R_{x\,(Standard)}} \cdot 1000 \qquad (2.1)$$

where, R_x denotes the ratio of rarer to the common isotope ($^{18}O/^{16}O$ or D/H). The δ notation has better resolution over the absolute ratios, facilitates intercomparison of results and aid in use of multi-component mixing models.

2.3 Isotopic Fractionation

The variation in the global distribution of various isotopes in different hydrologic systems is mainly brought about by isotopic fractionation. The isotopic fractionation, which is proportional to the differences in the mass of the isotope water species, may be described as the partitioning of the isotopes by physical or chemical processes. The chemical isotopic fractionation processes involve redistribution of isotopes of oxygen or hydrogen among different phases. The process can be either equilibrium isotopic reactions with equal reaction rates for both forward and backward reactions or non-equilibrium (kinetic) isotopic reactions with mass dependent reaction rates, which are unidirectional.

The distribution of isotopes between different phases of water, which are in equilibrium, is not uniform. In other words, the water vapour in equilibrium with liquid water is slightly depleted in heavier isotopes, in comparison to the liquid water. This is because of the differences in the hydrogen bond energy in the liquid phase between the isotopic molecules, resulting in the differences in the vapour pressure of the isotopic species of water. The difference in the ratio of heavier to lighter isotopes of one phase to that of the other is then defined by equilibrium fractionation factor α^+:

$$\alpha^+ = \frac{(R_x)_{liquid}}{(R_x)_{vapour}}$$

where, R_x is the ratio of concentrations of heavier to lighter isotopes.

The equilibrium fractionation factor has been determined by several investigators by vapour pressure measurement of pure isotopic species (142), by isotopic analyses of water and water vapour in equilibrium (95, 18, 77), or by dynamic distillation methods (17). However, the the following equation is widely used in hydrological studies for α^+ (95):

$$\ln \alpha^+ = AT^{-2} + BT^{-1} + C \tag{2.2}$$

where, T is the absolute air temperature in degree Kelvin, and A, B, C are coefficients. The values of the coefficients corresponding to $^{18}O/^{16}O$ are 1137, -0.4156 and -0.00207 respectively; and corresponding to D/H are 24844, -76.248 and 0.05261 respectively (95).

To study the effect of evaporation in the surface water bodies such as lakes, the fractionation factor, α^* is used in place of α^+. α^* is inverse of α^+. At the temperature of interest to a hydrologist the value of α^* is always less than one. The equilibrium enrichment factor, ϵ^* is more convenient to express the relative changes in the isotopic ratios of liquid water and water vapour in equilibrium. ϵ^* is expressed as $(1 - \alpha^*) * 1000$ (63).

Under natural conditions, the actual isotopic composition of water vapour is significantly lower than the values predicted using equilibrium enrichment factors. The difference between the total enrichment factor (ϵ) and the equilibrium enrichment factor (ϵ^*) is called as excess separation factor or kinetic enrichment factor ($\Delta\epsilon$). The kinetic enrichment factor is defined as follows (28):

$$\Delta\epsilon = (1 - h)\left(\frac{\rho_i}{\rho} - 1\right) \tag{2.3}$$

where, h is the relative humidity normalised to the liquid surface temperature, and ρ_i and ρ are the transport resistance of the rarer and common isotopic water species in air. Attempts to evaluate the kinetic enrichment factor have been made, through field evaporation pan experiments (57), through laboratory experiments (100) and through wind tunnel experiments (155). The values of the term $(\rho_i/\rho - 1)$, for oxygen and hydrogen isotopes have been proposed as 14.2 and 12.5 respectively (63).

A few investigators, who have made use of deuterium isotopes in lake studies, contended that in case of deuterium, the kinetic enrichment effect is small and it facilitates in avoiding complications that arise due to estimation of mean relative humidity (164, 165). However, it was shown that the kinetic effects in case of deuterium differs very much in field conditions from that in laboratory controlled experiments (167), supporting the earlier findings on the basis of evaporation pan experiments in lake Tiberias (57).

2.4 Isotopic Composition in Precipitation (δ_p)

The isotopic composition of atmospheric moisture and consequently precipitation, exhibits a broad spectrum of spatial and temporal variation. Subsequent to the initial attempts by several investigators (31, 32, 47, 53) to study the natural abundance of ^{18}O and D in meteoric waters, the available information on the global isotopic characteristics in fresh waters was summarised (27). Based on the IAEA/WMO Global Network for Isotopes in Precipitation (GNIP) database on monthly composite samples of precipitation, the following characteristics were identified (33, 162, 125).

Latitude effect: The heavier isotopic content decreases with increasing latitude. This is mainly because of the fact that the major global source of vapour is the tropical oceans i.e. between 30°N and 30°S latitudes. The poleward transport of this vapour results in the rain-out process and consequent depletion in the isotopic ratio of precipitated water.

Continental effect: The heavier isotope ratio decreases inland from the coast (approximately 2‰ per 1000 km). Basically, the rain-out process is responsible for this effect, implying that oceans are the major sources of vapour that precipitates over the continents. However, studies over Europe, South America and Northern India reveal that substantial amount of evapo-transpirated water from the plant covers returns to the atmosphere (115, 61, 85).

Seasonal effect: Winter precipitations are depleted in ^{18}O and D relative to the summer precipitations. This effect is more pronounced in mid and high latitude regions, due to seasonal variation in total precipitable water, seasonally modulated evapotranspiration over continental regions and seasonally changing source areas of the vapour and different storm trajectories (125).

The amount effect: An apparent relationship between the amount of rainfall and its isotopic ratio exists, i.e. for greater amount of rainfall, the isotopic ratios are more depleted (33). This effect is more pronounced during periods of low precipitation, because of the evaporative enrichment of raindrops as a consequence of lower humidity. Conversely, high intensity and heavy rainfall tend to modify the isotopic ratio of the surrounding atmospheric water vapour beneath the cloud base through exchange processes, thereby preserving the in-cloud isotopic signatures.

Altitude effect: When saturated air moves upward, it cools, resulting in condensation and consequent release of heat, which in turn counteracts the cooling. The fractionation of stable isotope takes place during this process. The adiabatic lapse rate, though it varies with altitude, is about 0.6°C/100m. For $\delta^{18}O$, the temperature dependence during adiabatic cooling is about 0.5‰/°C (33). This means that there may be about 0.3‰ variation in $\delta^{18}O$ per 100 metres variation in altitude. The altitude gradient (per 100 metres) varies between -0.1 ‰ and -0.5 ‰ for $\delta^{18}O$ and -2.5 ‰ and -4 ‰ for δD (23).

The adiabatic lapse rate of the cloud mass and hence the altitude effect on stable isotope ratios varies from region to region, due to variations in local topography. The altitude effect is further enhanced during the fall of raindrops, as it can evaporate with greater evaporation for raindrops that travel longer vertical distances (25). However, often the altitude effects are not observed for rainfall in the lee side of the mountainous terrain and also in case of precipitation in the form of snowfall.

Relationship with geographical / climatological parameters: To derive a relationship between the mean isotopic composition in the precipitation (dependent variable) and different basin geographical and climatological parameters (as independent variables), multiple linear regression analyses were carried out and an equation of the following type may be used (162):

$$\delta^{18}O = a_0 + a_1T + a_2P + a_3L + a_4A$$

where T, P, L and A are the monthly mean temperature (°C), monthly precipitation (mm), latitude (degrees) and altitude (m above m.s.l.) respectively; and a_0, a_1, a_2, a_3 and a_4 are regression coefficients. However, the proponents cautioned that the findings are true only on a global scale and suggested that on a regional scale, either amount effect or evaporation effect may be equally important in determining the spatial variability.

$\delta^{18}O$ - δD relationships: The relationship between $\delta^{18}O$ and δD in freshwaters was noted as early as 1953 (53). A best fit line was proposed (Craig's line) for the $\delta^{18}O$ and δD data of samples collected from different parts of the world (27):

$$\delta D = 8\,\delta^{18}O + 10 \qquad (2.4)$$

Currently, the equation of the Global Meteoric Water Line (GMWL) developed using the long-term weighted mean hydrogen and oxygen isotopic ratios collected under GNIP database is (125):

$$\delta D = 8.2\ \delta^{18}O + 11.27 \tag{2.5}$$

The above equation is identical to the earlier equations of the GMWL (33, 162) and also confirms that Craig's equation is good approximation of the points representing average isotopic composition of global freshwaters. However, the relationship between $\delta^{18}O$ and δD as described by Equation (2.5) is only apparent and as such cannot be used to derive the isotopic ratio of hydrogen with given isotopic ratio of oxygen for any location. This is because the GMWL is essentially an average of several Local Meteoric Water Lines (LMWL). Deviations in the $\delta^{18}O$ - δD relationship at different stations may be due to differences in the initial vapour isotopic composition, initial dew point temperature, extent and way of cooling, and kinetic effects during fall of rain drops (33).

The slope of the Regional MWL for Indian monsoon, proposed by (16) as 7.2 on the basis of extensive groundwater samples collected in the northern India, is distinctly different from that determined as 8.4, on the basis of long-term annual weighted mean values of isotope ratios in the precipitation samples, collected at New Delhi (36). The variation in the slope of the MWL indicates that the relation between $\delta^{18}O$ and δD is more complex in case of Indian monsoon precipitation.

'd' excess parameter The 'd' excess parameter or d-index means the surplus deuterium relative to the Craig's Line (33). The characteristics of the d-index are:

a) equilibrium processes do not change the d-index for any of the phases;

b) non-equilibrium evaporation from a limited amount of water reduces the d-index of the water as long as the exchange is not a dominating factor;

c) for non-equilibrium evaporation from an infinitely large and well mixed reservoir, d-index of the water will remain constant and that of the vapour will be positive and increase with the rate of reaction; and

d) the average d-index of precipitation at a given locality reflects the rate of evaporation in the source area.

2.5 Lake - Groundwater Interaction Studies

A lake generally acquires a specific isotopic labelling due to the process of evaporation and therefore, is significantly different from the adjacent groundwater system recharged by local precipitation. The lake is comparatively enriched in heavier isotopes. Since evaporation of the lake does not always takes place under equilibrium conditions, the lake is said to be 'doubly labelled' as the d-index of the lake will also be different compared to the groundwater. Although, evolution of the isotopic composition of a lake depends upon several processes such as channelised and sheet form surface inflow, sub-surface inflow from porous or fractured aquifers, direct precipitation etc., it is chiefly controlled by open surface evaporation process.

2.5.1 Lake Evaporation

The mechanism of lake evaporation is best explained by the Craig and Gordon linear resistance model (28):

a. Release of water vapour in isotopic equilibrium with liquid phase. A saturated sub-layer at the water-air interface which, compared to the liquid phase, is depleted in the heavier isotope by ϵ^+ (58).

b. Migration of this vapour away from the interface through a zone where, its transport is governed by molecular diffusion. This results in a further depletion of heavier isotopes.

c. The vapour reaches a fully turbulent region with no further fractionation and mixes with the existing vapour.

d. Vapour from the turbulent region may also penetrate through the diffusion layer and condense on the liquid surface. This process is termed as molecular exchange of the liquid with atmospheric vapour.

The simplified Craig and Gordon evaporation model can be written as:

$$\delta_E = \frac{(\alpha^{*}\delta_L - h\delta_a - \epsilon)}{(1-h+10^{-3}\Delta\epsilon)} \tag{2.6}$$

where, δ_L is the isotopic composition of the lake surface, δ_a is the isotopic content of the atmospheric water vapour, h is the relative humidity normalised to temperature at the interface and α^*, ϵ, $\Delta\epsilon$ are as given in section 2.3.

Suitable ways to determine or by-pass δ_a have been attempted by several investigators. Different methods are discussed below.

Direct sampling Samples of atmospheric water vapour can be collected, without fractionating the isotopes, from a suitable place near the lake and be analysed for the isotopic composition. Samples should be collected at the point where relative humidity is measured, regardless of its position in relation to the lake as the water vapour produced from the lake has little influence on the isotopic composition of the net vapour flux from the lake (39). In a study on Sparkling lake, Wisconsin, air samples from lake's windward shore were collected by pumping 1.5 lpm of air through a liquid nitrogen trap (82). This method of collection has also been outlined by many investigators. A description of such samplers has been presented (163). In the study on lake Tiberias, cold traps were used to collect atmospheric water vapour for δ_a determination (57). An automatic water vapour sampler was used during the study of Waidsee and Wiesensee lakes of Germany (164).

Due to low vapour pressure at high altitudes, it may sometimes be difficult to use cold traps to collect samples of atmospheric vapour in sufficient quantities required for isotope analyses. Hence, automated or continuously operatable mechanical devices are better suited for the purpose, however cost could be a constraining factor.

Index lake method The method was first used in a deep terminal lake with constant isotopic composition (39). The assumptions, that were made by the investigator in adopting the method, are: a) region of uniform climatic characteristics, b) no significant aerial variation in isotopic composition of atmospheric water vapour, c) surface temperatures of the lakes are close to the atmospheric temperature, and d) relative humidity has no large variation.

Since the lake is terminal type, that is the loss of water is only through evaporation, the following relationship holds:

$$\delta_E = \delta_I \tag{2.7}$$

For a terminal lake which has reached both hydrologic and isotopic steady state, Equation (2.7) could be substituted in Equation (2.6) and δ_a be determined, which can then be subsequently used for other nearby lakes.

This method requires the existence of an index lake, which meets all the aforesaid assumptions, near the lake(s) to be investigated. This may not be feasible in most cases, but if such a lake is available, this could be the most suitable method.

Evaporation pan method For an isolated water body, such as evaporation pan, the following expression is valid (157):

$$(\delta - \delta_s)/(\delta_o - \delta_s) = f^m \tag{2.8}$$

and

$$m = (h - \epsilon)/(1 - h + \Delta\epsilon) \tag{2.9}$$

where,
- f = fraction of volume of water remaining at time t
- δ = isotopic composition of water
- δ_s = steady-state isotopic composition of water
- δ_o = initial isotopic composition of water

Equation (2.8) can be rewritten as,

$$\ln(\delta - \delta_s) - \ln(\delta_o - \delta_s) = m \cdot \ln f \tag{2.10}$$

According to Equation (2.10), slope of the best fit curve of ln (δ - δ_s) versus ln (f) will be m_{pan}, whereas, δ_s can be determined on trial basis.

If the mean lake surface temperature and pan temperature do not vary much, then by modifying Equation (2.6) and assuming negligible error in applying m and δ_s of a pan to the lake, δ_E can be determined by using the following equation:

$$m_{pan} = [(\delta_E)_{lake} - \delta_{lake}]/[\delta_{lake} - (\delta_s)_{pan}] \qquad (2.11)$$

On similar lines, a leaky evaporation pan theory was used to investigate the water balance of a reservoir in Brazil (140).

However, as the pan evaporation seldom proceeds under steady environment conditions to facilitate these computations, a modified version of this approach was proposed (3).

Multiple pan method The effect of pan environment, on the parameters, which influence the value of δ_E, have been studied (3). The researchers, measured the evaporation and isotope ratios from four evaporation pans placed under different conditions. They observed that the δD enrichment rarely proceeded in a uniform manner and sharp changes in relative humidity, and not precipitation, influenced the breaks in δD - f curves. They recommended the use of a two-pan method, with one pan filled with water of spiked isotope concentration, to avoid dependency on relative humidity values, but this was not supported by field experiments. Further, they stated that location of the pan is not an important criteria under normal conditions. The pan methods are yet to prove their applicability in different environments. But these methods may be used if relative humidity does not show a large variability.

δ_p equilibrium assumption method The relationship between isotopic composition of atmospheric water vapour and precipitation has long been recognised (166). The isotopic composition of continental water vapour, on rainy days, is in equilibrium with that of precipitation (29). Assumption that δ_a is in equilibrium with the mean isotopic content of precipitation is valid (167) and therefore,

$$\delta_a = \alpha' \delta_p - \epsilon' \qquad (2.12)$$

The study on Sparkling lake supported the above assumption (82). The measured δ_a values compare well with those computed using Equation (2.12) for measured relative humidity and air temperature. The deviations (about 2.5‰) between observed and calculated δ_a, for the warmer months during the study period, were possibly due to contributions from numerous lakes situated near the lake under consideration (82). Inspite of few constraints, this method seems to be reasonable if short-term δ_p is available.

2.5.1.1 Dissolved salt effect

The concentration of the dissolved salts in lake water influences the isotopic composition by three ways under evaporating conditions. Firstly, the dissolved salts decrease the thermodynamic activity of the water as well as the evaporation rate. Secondly, the water molecule entering the hydration sphere of certain ions causes isotopic fractionation compared to free water. For example, in the case of oxygen isotopes, there is a notable difference between CO_2 equilibrated with pure water and that with a salt solution of the same water. The values of hydration fractionation factors of some electrolytes have been presented (58). A third salt effect, though minor, is due to the crystallisation of salts. Some deposited salts contain water of crystallisation which generally has an isotopic composition different from that of the mother water. The above effects become very important for saline lakes, sebkhas (ephemeral lakes) and for drying up lakes in the last stages of evaporation. However, the salt effect on evaporating water bodies can be neglected in most natural cases (63).

2.5.2 Quantitative Evaluation of Lake - Groundwater Interaction

The lake water - groundwater interaction can be quantitatively evaluated by focusing on the lake system and using the mass balance approach. One of the basic assumptions in the mass balance approach is that the lake is well mixed with uniform isotopic composition. However, there are lakes that do not confirm to this assumption. Poorly mixed lake have been discussed under section 2.5.2.2.

2.5.2.1 Well mixed lakes

The basic isotope mass balance equation can be simplified by taking the type of lake and lake environment into consideration. A lake that gains substantial quantity of water from the adjacent aquifer is called as "discharge lake" or "seepage lake", and a lake that loses water to adjacent aquifer is called as "recharge lake" or "groundwater lake". Lakes that lose water only through evaporation are called as "terminal lakes". There are variations to these definitions if a lake is having no inflow or inflow compensates the evaporation. A lake that has substantial subsurface inflow and subsurface outflow is called as a "flow-through lake". A range of lake conditions and the equations which describe evolution of the isotopic composition of lakes and that can be used to compute the isotopic mass balance, has earlier been reviewed in detail (63). Some lake types have been presented below:

Case I: Flow-through lakes

The lake and groundwater interaction can be assessed, by simultaneously solving the water balance and isotope mass balance equations. The conventional water balance equation for the lake may be written as:

$$\Delta V = S_I + P_I + SS_I - S_O - E_O - SS_O \tag{2.13}$$

where, ΔV is the change in volume, S_I surface inflow, P_I is direct precipitation on lake, SS_I is groundwater inflow, S_O is surface outflow, E_O is evaporation from lake surface and SS_O is seepage from the lake. Established conventional techniques can be used to measure or estimate all the above components within certain error limits (158) except the sub-surface ones (SS_I and SS_O). Although use of specially designed seepage meters have been reported (92), the variation of subsurface flow in time and space and the practical problems of installation and periodical sampling in deep lakes, restricts their use in most cases.

The isotope mass-balance equation for the lake is given by

$$\frac{d\,(\delta_L V)}{dt} = \delta_s\, S_I + \delta_p\, P_I + \delta_g\, SS_I - \delta_L\, S_O - \delta_E\, E_O - \delta_l\, SS_O \tag{2.14}$$

where, t is the time for which the balance is being computed, other notations with subscripts I and O are as per Equation (2.13), and δ-notations are the corresponding isotope ratios. The equation has been simplified by assuming δ_L to be equal to $\delta_{S\text{-}O}$ and $\delta_{SS\text{-}O}$ and δ_g to be equal to $\delta_{SS\text{-}I}$. In Equation (2.14), except δ_E all other components are directly measurable.

The following equation was proposed to estimate the outflow to evaporation ratio, which can be used with conventional data to estimate the subsurface outflow components (167):

$$x = \frac{O}{E_O} = \frac{\left(\frac{S_I}{E_O}\right)(\delta_g - \delta_S) + \left(\frac{P_I}{E_O}\right)(\delta_g - \delta_P) + \delta_E - \delta_g}{(\delta_g - \delta_L)} \qquad (2.15)$$

Equation (2.15) may further be simplified by assuming δ_g to be equal to δ_P, in absence of the groundwater data (167). Using the Equation (2.15) the subsurface outflow from Lake Chala, Tanzania was calculated using oxygen isotope ratios as 3.9 * 10^6 m^3/yr (167), which was in very good agreement with the value of 4.14 * 10^6 m^3/yr, determined by injected artificial tritium for whole body tracing (110).

Equation (2.15) was also used in the study of Lake Bagry, Poland to estimate the subsurface inflow and outflow (167). Lake Bagry did not have any surface inflow or surface outflow. The investigator used the long term averages of precipitation, temperature, humidity and evaporation data of the study area. Though oxygen isotopic composition of the lake showed seasonal variation, the mean value was estimated by integrating the area encompassing δ_l - time curve. The residence time (volume / annual outflow) calculated by using the results of isotope method compared well with that obtained through electrical analogy approach, which confirmed validity of the isotope approach (167).

The groundwater exchange with Lake Sparkling, Wisconsin, USA was estimated using the stable isotope mass balance approach with oxygen isotopes (82). Since the surface inflow to the lake was insignificant and the lake was in an isotopically steady state condition, using Equations (2.13) and (2.14) the investigators derived:

$$SS_I = \frac{P_I(\delta_L - \delta_P) + E_O(\delta_E - \delta_l)}{\delta_g - \delta_l} \qquad (2.16)$$

The above equation facilitated them to determine the groundwater inflow independent of groundwater outflow rate. The results showed that the subsurface flow accounted for about 27% of the total inflow and 50% of the total outflow. The results were found comparable with those obtained by a three dimensional flow and solute transport model (83) for the lake - aquifer system.

The Sparkling lake study was further extended, considering the lake to be an Index lake, in order to study the groundwater components of other nearby lakes with negligible surface components (84). The investigators studied the groundwater exchange of Lake Crystal, Lake Pallette, Lake Big Musky. Assuming isotopic steady state of the lake, they used the groundwater inflow data of Sparkling lake to evaluate the subsurface components of the three lakes. The success of this approach depends largely on accuracy of estimation of the water balance components of the Index lake.

The groundwater component in the water balance of Williams lake, Minnesota, USA was estimated using a multi-tracer approach and flow net approach (91). The investigators found a large difference between the results obtained through isotope approach and flow net approach. It was contended, that the discrepancy was mainly due to the errors involved in the estimation of evaporation losses and the isotopic composition of the lake evaporates through indirect methods. The investigators reported that both the values obtained for groundwater inflow and the range of uncertainty varied with the selection procedure of the groundwater isotopic index. There was a difference of approximately 2‰ between the isotopic composition of groundwater and the weighted precipitation index. They also considered two different values of lake evaporates as input to the water balance equations. The investigators did not, however, consider the selective recharge process in the study area. The results showed that the oxygen isotope was a better tracer in accounting the exchange of water of lake Williams with groundwater than the four chemical tracers viz. chlorine, sodium, magnesium or dissolved organic carbon. Therefore, the combined use of hydrogeological and chemical approaches, improve the estimation of the lake - aquifer exchanges (91).

Case II: Desiccating terminal lakes with no inflow

For a terminal lake with no inflow, the variation of isotopic composition can be written as (63):

$$\frac{d\delta_L}{d\ln f} = \frac{h(\delta_L - \delta_a) - (\delta_L + 1)\left(\Delta\epsilon + \frac{\epsilon^*}{\alpha^*}\right)}{1 - h + \Delta\epsilon} \tag{2.17}$$

where, f is the fraction of remaining water. Assuming environmental conditions of evaporation as constant i.e. keeping h, δa, $\Delta\epsilon$ and ϵ^+ as constant, the Equation (2.17) can be integrated to get

$$\delta_L = \left(\delta_{L_0} - \frac{A}{B}\right) f^B + \frac{A}{B} \tag{2.18}$$

where, δ_{lo} is the initial lake isotopic composition, f is the fraction of water remaining, and A and B are defined as follows:

$$A = \frac{h\delta_a + \Delta\epsilon + \frac{\epsilon^*}{\alpha^*}}{1 - h + \Delta\epsilon} \tag{2.19}$$

$$B = \frac{h - \Delta\epsilon - \frac{\epsilon^*}{\alpha^*}}{1 - h + \Delta\epsilon} \tag{2.20}$$

From Equation (2.18) it is seen that as f tends to zero (i.e. during final stages of evaporation), the term (A/B) becomes the final isotopic composition of the lake and it is therefore independent of initial isotopic composition of lake.

Case III: Lakes with slow and unidirectional change in volume

For lakes and reservoirs with long residence time and slow and unidirectional change in volume, the following equation holds if we assume the isotopic composition of inflow waters and atmospheric air, the environmental conditions and the inflow, outflow and evaporation rates to be constant (63):

$$\delta_L = \left(\delta_{l_o} - \frac{\delta_L + AX}{1 + BX}\right) f^{\frac{-(1 + BX)}{(1 - X - Y)}} + \frac{\delta_L + AX}{1 + BX} \tag{2.21}$$

where, X = Evaporation / Inflow and Y = Outflow / Inflow.

Case IV: Lakes of leaky evaporation pan type

This type of lake is similar to Case III, but with no inflow i.e. I = zero.

$$\delta_L = \left(\delta_{L_0} - \frac{A}{B}\right) f^{z} + \frac{A}{B} \qquad (2.22)$$

where, Z = E / (E + Q), that is the ratio of evaporation to total losses from the lake.

For the water balance study of Quebra Unhas reservoir in Brazil, this approach was used (140). Since there was negligible surface inflow and no surface outflow from the reservoir, the investigators used the leaking evaporation pan analogy to estimate the evaporation and subsurface outflow from the reservoir. Equation (2.22) was used with the oxygen isotope data of the pan to evaluate the values of parameters A and B, which were again used in Equation (2.22) to evaluate the parameter Z of the reservoir. The investigators estimated that around 67% of the total losses from reservoir were accounted by evaporation and around 33% by subsurface outflow. An independent tracer, viz. chloride ion was also used for the study which indicated that subsurface outflow accounted for around 25% of total losses. As per the investigators, the results of two independent tracers can be said to be in reasonable agreement if the errors associated in the computations are taken into consideration.

Case V: Lakes with constant volume

For lakes with constant volume i.e. dv/dt = 0, we have I = Q + E. Assuming E/I = X and Q/I = 1 - X, where, X is the fraction of water lost by evaporation and V/I = T, where, T is the mean residence time of the lake, we get:

$$\frac{\delta_L}{\delta_t} = -\frac{1}{T}[(1 + BX)\,\delta_L - \delta_I - AX] \qquad (2.23)$$

Cconsidering evaporation conditions and the isotopic composition of inflow water to be constant and integrating Equation (2.23), we get:

$$\delta_L = \left(\frac{\delta_I + AX}{1 + BX}\right) + \left(\delta_{L_o} - \frac{\delta_I + AX}{1 + BX}\right) e^{-(1 + BX)\frac{t}{T}} \tag{2.24}$$

As per the above equation, when t tends to infinity, δ_l tends to δ_s and a steady state value is achieved, which is defined by:

$$\delta_s = \frac{\delta_I + AX}{1 + BX} \tag{2.25}$$

The groundwater components in the water balance of two young artificial lakes in Germany viz. Lake Wiesensee and Lake Waidsee were investigated by using the above equations (164). The lakes had no surface inflow or outflow. The investigator considered the lakes to be of constant volume, although there was a 6% decrease in volume. During the investigation period from 1970 to 1974, the lakes showed a continuous enrichment in the heavy isotope (δD) due to evaporation and seasonal variations in the precipitation. The investigator correlated ln(δ_l^S - δ_l) with time (t) and selected the value of δ_l^S by trial using Equation (2.24). The values of parameters A and B were computed using the field humidity and δ_a values. The results pertaining to the estimated evaporation and subsurface inflow and outflow were found to be of the order of 30%. The investigator concluded that the stable isotope method is suitable for computing subsurface components because there is no other better method to evaluate these components with comparable accuracy. Using Equation (2.15), the data of Lake Waidsee was reinterpreted (167) and based on discrepancy in the X value, it was shown that the results of the earlier investigation (164) are questionable.

Case VI: Terminal lakes where evaporation compensates inflow

When the evaporation loss from a lake exactly compensates the inflow to the lake, then the steady state isotopic composition of the lake can be determined as (63):

$$\delta_S^T = \alpha^{\cdot}\delta_I(1 - h + \Delta\epsilon) + \alpha^{\cdot}h\delta_a + \alpha^{\cdot}\Delta\epsilon + \epsilon^{\cdot} \tag{2.26}$$

δ_S^T is the maximum enrichment in heavy isotope in a constant volume lake. Correction terms may have to be included in Equation (2.26) as in certain terminal lakes, the salt effects may dominate.

A terminal type Lake Burdur, Turkey was used as an index lake, for molecular exchange in the region where two fresh water lakes viz. Lake Egridir and Lake Beysehir were situated (39). He used the following equation to estimate the water balance of the fresh water lakes:

$$\frac{d(\delta_L V)}{dt} = I(\delta_i - \delta_l) - E\left(\frac{\delta_L - \delta_L^T}{\alpha\,(1-h)} + \delta_l^T - \delta_l\right) \tag{2.27}$$

where, I is the inflow rate, E is the evaporation rate, δ is the isotopic composition and subscripts i and l denote inflow and mean lake values respectively. The superscript T denotes the data pertaining to terminal lake (index lake). The investigator stressed the sensitive nature of the isotopic composition of lake water in using the index lake method. An error of 1% in estimated mean isotopic composition resulted in 5% change in the inflow. However, the method was found less sensitive to the fractionation factors, relative humidity and kinetic effects.

2.5.2.2 Poorly mixed lakes

The study of water balance of a poorly mixed lake may be relatively difficult. However, with a dense network of sampling and properly weighted mean values, attempts may be made to solve the mass balance equation with a reasonable degree of accuracy. The systematic variation in the isotopic composition of a shallow lake was utilised, disregarding mixing and considering each sample as representative of different stages of evaporation of the lake (51). While studying a large swamp area, the investigators combined the isotopic and water balance equations with the salt balance equations (40). They suggested that it allows the calculation of the ratio of evaporation to evapotranspiration.

The water balance of lake Bracciano, Italy has been investigated using isotope approach (64). They suggested that 90% of the error in the calculated parameter X, was due to the errors associated with humidity and isotopic composition of the atmospheric air and it is difficult to measure or estimate these parameters with reasonable accuracy. Hydrogen isotope ratio was used to study the subsurface components of Lake Neuseidl, Austria, which is a shallow lake with poor mixing (165). The investigators neglected the kinetic enrichment effect in their computations. It was concluded that large errors associated with δ_E, resulted in the determination of evaporation rate not better than 50% accuracy.

The seasonal stratification is not uncommon in temperate lakes, and tropical or subtropical mountainous lakes and permanent stratification in deep lakes. Under stratified conditions, the upper portion of the lake, called epilimnion, is enriched in heavier isotopes (due to evaporation) compared to the bottom waters, called hypolimnion. In the water balance study of Lake Tiberias, which is a stratified lake, an approach to include the effect of isotope exchange between the epilimnion and hypolimnion (eddy diffusion) and also the increase in the volume of the epilimnion (hypolimnion entrainment) due to the subsidence of thermocline, was employed (57). The effect due to exchange with hypolimnion is usually not significant, if it is the only process affecting the isotopic composition of the hypolimnion. However, the effect of hypolimnion entrainment is generally significant (60).

2.5.3 Qualitative Studies on Lake - Groundwater Interaction

Isotope methods are also useful in evaluating the hypothesis of lake water interaction with local groundwater or springs emerging at greater distances from the lakes. The stable isotope technique was used to find the direction of subsurface outflow from a lake in Germany (139). The belief that few springs in the vicinity of lake Chala, Kenya are fed by the lake, was disproved by isotopic investigations (109). On the basis of stable isotope data, the investigator showed that the lake's contribution to the springs is negligible. Recently, oxygen isotope ratios have been used to establish the connection between the Mt. Galunggung crater lake, Indonesia and a nearby spring (1).

2.6 River - Groundwater Interaction Studies

The spatial and temporal variations in the isotopic composition of river waters (δ_R) are mainly influenced by the number and type of its sources, and to some extent by evaporation from the river surface in certain climatic regions. Variations in the observed δ_R, reflect the variable contributions from isotopically different sources, and can be evaluated if isotopic indexes of the sources are known. However, the river water isotopic characterisation and its utility in studying river - aquifer interactions depends greatly on the spatial and temporal scale of the problem.

Amplitude of seasonal (temporal) variations in δ_R, gives some insight into the hydrologic regimes that control the flow in a river. Seasonal variation of the isotopic composition of river water, will be more pronounced in case of rivers that are dominated by surface runoff than those dominated by groundwater inputs. This is due to the fact that groundwater bodies are isotopically constant and closely reflect the average annual isotopic composition of local precipitation (even if seasonal variations in the precipitation isotopic composition are large). The rivers fed by karstic systems with a short residence time (τ) will show a pronounced seasonal variation in δ_R, than the ones fed by porous aquifers (where τ is longer). On the other hand, in precipitation dominated basins, the seasonal variation in precipitation isotopic composition is reflected in more enriched δ_R during summer and depleted δ_R during winter. But snowmelt during summer may reverse the above trend or in some cases shift the trend. The isotopic characteristics of river water (both precipitation dominated and groundwater dominated) are also affected by the altitude effect in precipitation and reflected in both surface and sub-surface components of the river flow. It is also noticeable that the altitude effect might produce changes in δ_R which over-impose changes due to the seasonal variation of δ_P.

In case of small drainage basins, the groundwater reservoir may be limited and the river discharge is largely dependent on the precipitation events. In case of some alpine basins, the magnitude of seasonal variations in isotopic composition vary inversely with the size of the basins. In mountainous terrains, the evaluation of the variable contribution from different sources using the isotopic characteristics of the sources can be rather difficult because of rapid release from surface and sub-surface reservoirs to the river flow.

2.6.1 Two-Component Mixing Model

The flow in a river at any point is a mixture of two components viz., the base flow integrated upto the location of interest and the runoff generated in response to a precipitation event. If these two are isotopically different, the isotopic composition of the stream water at the point of interest is changed. The extent of change is a function of proportion of runoff to the baseflow. This can be expressed by the following equations which conform to the law of mass conservation:

$$m_r = m_1 + m_2 \tag{2.28}$$

$$m_r C_r = m_1 C_1 + m_2 C_2 \tag{2.29}$$

where, m is the quantity of flow components expressed in fraction, C is the tracer concentration, subscript r denotes stream admixture at the point of interest, and the subscripts 1 and 2 denote the two components contributing to the stream. If stream discharge data is available at the sampling point, then the quantity of components could be expressed in terms of volume or depth. In absence of the discharge data, m_r could be assumed as one and the m_1 and m_2 could be expressed as ratio to the total discharge at a particular time. Rewriting Equation (2.28), we get:

$$m_1 = 1 - m_2 \tag{2.30}$$

Substituting Equation (2.30) in (2.29) and rearranging, we get:

$$m_2 = \frac{C_1 - C_r}{C_1 - C_2} \tag{2.31}$$

Equations (2.28) and (2.31) could be used to compute fractions of the two components of stream flow at a given point in space and time. However, the two-component mixing model can be successfully used for hydrograph separation only if the following assumptions are valid (137):

a) the isotopic composition of the two end members be unique or the variations known and well documented;

b) the isotopic composition of the end members are significantly different and the difference is larger than the mass spectrometer analytical precision; and

c) the contributions from the vadose water and from surface detentions or storage are negligible during investigation.

2.6.2 Three-Component Mixing Model

In certain cases, the assumption that there are only two components may not be valid, and the isotopic value of the admixture may fall outside the mixing line defined by the two suspected end members (80,38). The third component in the above cases is often found to be soil water or in some cases channel precipitation. The effect of a summer rainfall event on the stream flow process was studied in a small Pennsylvania hill watershed (117). The investigators, found that during a single storm event, the near stream groundwater table rises without substantially affecting the regional water table; and extend the associated subsurface discharge and surface runoff zones. In this hydrologic process that affect streamflow volume and quality, three components viz. channel precipitation, surface runoff and groundwater are initiated sequentially. Further, the dominant component of the stream flow also changes during the storm event (117).

Separation of the hydrograph of stream generated by three components can be carried out using the following equation (38), provided contribution of one of the components is known:

$$\frac{Q_1}{Q_r} = \left(\frac{C_r - C_3}{C_1 - C_3}\right) - \frac{Q_2}{Q_r}\left(\frac{C_2 - C_3}{C_1 - C_3}\right) \tag{2.32}$$

where, Q is the discharge (in units of m^3 or m), C is the tracer concentration and subscript r refers to the admixture or stream at the sampling point, subscripts 1, 2 and 3 represent the three end members contributing to the flow.

In absence of quantitative information on the contribution of any of the components, trilinear graphical technique can be adopted. $\delta^{18}O$ and SiO_2 data pertaining to the three different flow components in a microwatershed located in the Harp lake catchment, Ontario, Canada were utilised by researchers (69). They proposed the following set of equations that can be solved to determine the relative contribution of the three components of the stream flow, if two tracers are used:

$$\frac{Q_1}{Q_r} = X_n = \frac{[(C^b_r - C^b_2)(C^a_3 - C^a_2) - (C^a_r - C^a_2)(C^b_3 - C^b_2)]}{[(C^b_1 - C^b_2)(C^a_3 - C^a_2) - (C^a_1 - C^a_2)(C^b_3 - C^b_2)]} \tag{2.33}$$

$$\frac{Q_2}{Q_r} = \left(\frac{C^a_r - C^a_3}{C^a_2 - C^a_3}\right) - X_n\left(\frac{C^a_1 - C^a_3}{C^a_2 - C^a_3}\right) \tag{2.34}$$

$$\frac{Q_3}{Q_r} = \left(\frac{C^a_r - C^a_2}{C^a_3 - C^a_2}\right) - X_n\left(\frac{C^a_1 - C^a_2}{C^a_3 - C^a_2}\right) \tag{2.35}$$

where, Q is the discharge (inunits of m^3 or m), C is the tracer concentration, superscripts a and b denote the two tracers used for the study, and subscripts 1,2,3 and r denote the three inflow components and the river, respectively. For a successful application of three component mixing model, the following assumptions should be valid (69):

a) there are only three sources that contribute to the streamflow;

b) isotopic and/or chemical concentration of the sources must be distinct;

c) the concentration of the three suspected sources should not be collinear for either of the tracers;

d) the concentration (or index) of each suspected source should be constant during the event being studied; and

e) the tracers mix conservatively.

As a constant value of the end members is an essential precondition for any of the models given above, it is imperative that the variations, if any, in the end members be known and well documented. The possible variations in the end members, as observed by various investigators based on the studies carried out in different climatic characteristics, are described below:

Runoff: The variation in the runoff component may be induced by temporal variations in the isotopic composition of local precipitation (99, 116). In forested catchments, the canopy intercepted rainfall may have a different isotopic composition from rainfall in open area (129). In case of large watersheds or catchments with significant altitude difference (altitude effect), the new water may show a large spatial variability. The infiltration characteristics and subsequent storage of precipitation in the soil zone at different time periods of rainfall, can be deciding factor in mountainous catchments. Variation in the isotopic composition of precipitation in three different types of forest viz. deciduous, spruce and pine was studied and compared the same with open area (37). The investigators reported that the $\delta^{18}O$ of precipitation in the forests were enriched by 0.1‰ to 0.2‰ compared to the open area rainfall. The maximum difference observed during their study period was about 1.5‰ during a particular storm in the spring season. The above facts, highlight the need for better spatial resolution of precipitation isotopic data in storm-scale stream flow studies.

Groundwater: In case of groundwater, the variable residence time and consequent heterogeneity in groundwater isotopic composition poses problems in arriving at a representative groundwater index. Methods like using the mean value from large number of samples or the skewed value of the frequency distribution are sometimes used. Use of pre-storm baseflow or interpolated interstorm baseflow as the appropriate groundwater index is also advocated (72).

Soil water: In most of the studies, investigators consider soil water and groundwater together as a single variable in the mixing models. Soil water accounted for about 50% of the streamflow in Mattole river basin, California (80). The researchers, used ^{18}O and D as tracers and found the streamflow to be a mixture of earlier rainfall and recent rainfall (at the time of their investigation). They also found that the earlier storm water was stored in soil zone which subsequently joined the streamflow. The variability in the soil water isotopic composition is caused by evaporation (62). Homogeneity of soil water isotopic composition also depends upon the soil type. The homogenisation is more pronounced in clayey soil than sandy soils (20). However, it was reported on the basis of an intensive study in Maimai catchment, New Zealand, that there was a definite trend in soil water isotopic composition in downslope and down profile directions and the soil water is poorly mixed on storm time-scales (99).

Inspite of several studies in a wide range of hydrogeologic, climatic and landuse environments in Australia and New Zealand, no substantial advances have been made in modelling and prediction of the storm-scale flow processes (113). However, the studies have resulted in increased conceptual understanding of the streamflow generation processes, revealing their complex nature.

2.6.3 Application in Large River Systems

The isotopic composition of a large river yields little information on the characteristics of local hydrogeology or meteorological events, but in general reflects the regional hydrogeology and major events since the scale of homogenisation of isotopic composition from different sources is fairly small (55). There have been only limited number of studies on the isotopic characteristics of large rivers such as river Amazon, river Murray (Australia), rivers Tarapaca and El Loa (Chile), and river Missouri-Mississippi (USA). The studies have given insights into the effect of evaporation, mixing of large rivers, presence of reservoirs on the river course and contribution from groundwater on the isotopic composition of river waters. Some of the salient features of the findings have been discussed below.

Tarapaca and El Loa rivers in northern Chile were studied for their isotopic characteristics (7). The study revealed that the river Tarapaca, which is a large river with a catchment area of 33,000 km^2, did not show any sign of evaporation effect on the isotopic composition. However, in the lower part of the river basin, evaporation plays only a minor role during the low flow conditions. The investigators suggested that the isotopic composition of river Tarapaca is controlled by head waters and δ_R is only slightly modified along the river course. On the other hand, in river Loa located in the nearby area, the isotopic composition was found to be controlled by input from the groundwater of various origins and also by tributary inflow that showed some effect of evaporation. The Conchin reservoir across the river El Loa also resulted in a small evaporative enrichment and modified isotopic characteristics of the river.

The most comprehensive study on the isotopic evolution of a large river, with particular emphasis on the effect of evaporation, has been carried out on river Murray, Australia (133). The investigators reported that the isotopic composition of the river has large systematic changes in both time and space, showing an enrichment of 25‰ in δD between headwater storage and downstream end of the main channel of the river during summer months. It was suggested that the large variation could be explained through irrigation induced evaporative enrichment of tributaries and evaporation losses from the water surface of river Murray. However, the influx of saline groundwater does not significantly change the isotopic composition of the river Murray due to its small contribution to the river flow.

In case of large tributaries joining a large river, the lateral mixing and homogenisation is not instantaneous. The Liard - McKenzie river system, the investigators (86) found on the basis of isotopic data from 10 cross-sections along the river course that the complete mixing of the main river was achieved at a distance of around 300km downstream from the confluence of the two rivers. The difference in the δD isotopic composition of the two rivers (-21.3‰ and -17.4‰) were used to investigate the extent of lateral mixing. A similar study on Amazon river, made use of the mixing factor to compute the relative contribution of two rivers viz. Solimoes and Negro to the Amazon river (97). The investigators reported that the mixing of two tributaries was found incomplete even after a distance of 120 km downstream from the confluence. Such inhomogeneties in the isotopic composition across the river may pose problems in studying the river - groundwater interrelationships.

The isotopic composition of the river may sometimes be significantly modified due to the effect of evaporation from river surface. The studies on effect of evaporation on the isotopic characteristics of large rivers have been fewer and it is possibly a dominating factor only in dry or arid conditions. Apart from the study on river Murray, such effects have been noted in the investigations of Sebkhas (50), where the investigators utilised the isotopic enrichment to evaluate the evaporative loss in the rivers before they reach the lakes. On the other hand, no effect of evaporation was observed in Amazon river (97). The evaporation effect on rivers is usually not important in moderate climates (55). However, in arid regions, seasonal isotopic variations in river are caused not by variations in the precipitation but by the evaporation in lower basins (where precipitation is comparatively less). Inspite of the contributions from the tributaries carrying isotopically enriched water due to evaporation in lower basins, the main river generally retains its headwater isotopic characteristics, as found in the Missouri-Mississippi study (54).

The isotopic characteristics of river Ganga from its origin to Patna, located around 1800 km downstream of the river's origin point was investigated (120). The investigators found that the river is progressively enriched in its isotopic composition along its course due to tributary inflow and probable evaporative loss from the river surface. A slight depletion in the δ_R values in the lower basin of the study area was ascribed to probable groundwater contribution. The contribution of groundwater to river Ganga during non-monsoon months results in the progressive enrichment of river isotopic composition (105).

There are several methods for selecting an appropriate groundwater index. One of the method is to sample the groundwater, away from the influence of any surface water leakage, with no evidence of partial evaporation during infiltration process in the area and should have comparatively more positive values (111). Such a method was adopted in a study on river Ganga (105). In a study on Izanzo river, Italy (102), the more skewed value in the frequency distribution analyses of the isotopic composition of several groundwater samples, collected at increased distances from river, was identified as the groundwater index.

The river water isotopic characteristics can be used to monitor the extent of seepage from river to the adjoining aquifers. In northern Italy, the seepage of Reno river water into the groundwater system in response to increased withdrawal from the adjacent aquifers was studied by isotope techniques (22). The investigators incorporated the small difference of 1‰ in $\delta^{18}O$

and 5‰ in δD in the isotopic compositions of river and groundwater to estimate the extent of influence. The results were confirmed by environmental tritium data. The enriched isotopic compositions of river Colorado, due to a number of reservoirs across the river, were utilised to relate the increased salinity in the adjacent aquifers with differential seepage from the river (112). The isotopic investigations in the Gangetic basin lying in Bangladesh to study the surface water and groundwater interaction, revealed that the river Ganga does not contribute to the groundwater but the hydrometric data showed that the river determines the base level of groundwater and therefore the water table (43).

The isotope characteristics of Bishop creek in the Owens Valley, California, USA was studied (138). The investigators used difference in the δD values of the enriched upstream reservoirs and the local groundwater, snow cores and tributary inflow, to determine (and quantify) the locations of the groundwater inflow. They concluded that the groundwater inflow to the creek varied both in time and space, as reflected by the variable isotopic composition of the creek water along its course.

From the above review of literature, it has been observed that the isotope techniques are fairly well developed. They have been shown to be advantageous, in investigating surface water and groundwater interaction, in many countries world wide. But, in India, such studies on isotope hydrology are limited. The full potential of the isotope techniques, may be realised only upon trying them in different environments. In view of this, an attempt has been made to study the surface water and groundwater interaction in two different fresh water environments in India, by using the stable isotope techniques.

CHAPTER 3

STUDY OF LAKE NAINITAL AND GROUNDWATER INTERACTION

3.0 STUDY OF LAKE NAINITAL AND GROUNDWATER INTERACTION

3.1 Introduction

Lake Nainital (29°23'09" N and 79°27'35" E) is a high altitude (1937 m above m.s.l.) natural lake (Figure 3.1) located in Nainital district, Uttar Pradesh, India. It is a crescent-shaped lake with a maximum length of 1.4 km and width of 0.45 km. The maximum and mean depths of the lake are 27.3 m and 18.5 m respectively. The surface area of the lake is 0.46 km^2 with a maximum capacity of 8.57 Mm^3. The total population in the basin is around 40,000. The lake is a major summer resort in north India, and attracts nearly 100,000 tourists annually. Tourism is the major industry of this region and there is no agricultural or industrial activity within the lake catchment area.

3.1.1 Geology

The Nainital Massif occurs in-between the flat plains of the Ganga basin and the mature old terrain of the Lesser Himalaya, which stand out as a physiographic eminence to more than 2500 m height. It is defined as a mountain front of the Lesser Himalaya in Kumaun region. It forms the south-eastern extremity of the Krol belt that extends about 300 km from Solan in Himachal Pradesh. The Lesser Himalayan rocks and Nainital Massif (2000 - 500 m.a.), thrust over the Siwalik (20 - 2 m.a.) along the Main Boundary Thrust. This has resulted in severe compression of the Massif producing overturned folds, faults and fractures. Differential movements of fractured blocks have given rise to horst and graben structure that is described as block-faulted mountains (152).

3.1.1.1 Lithostratigraphy

The Nainital Massif is divided into lower Jaunsar Group and upper Mussoorie Group (150). The stratigraphic succession of the Mussoorie Group in the lake basin comprises of three formations viz. Blaini, Krol and Tal (Table 3.1).

Blaini Formation: The Blaini Formation is divided into two members - lower Pangot and upper Kailakhan units. The Lariakantha quartzite of Jaunsar Group is overlain by Pangot member without any structural discontinuity. Kailakhan Slate is best seen on the Kailakhan-Alukhet spur, south-east of Nainital town.

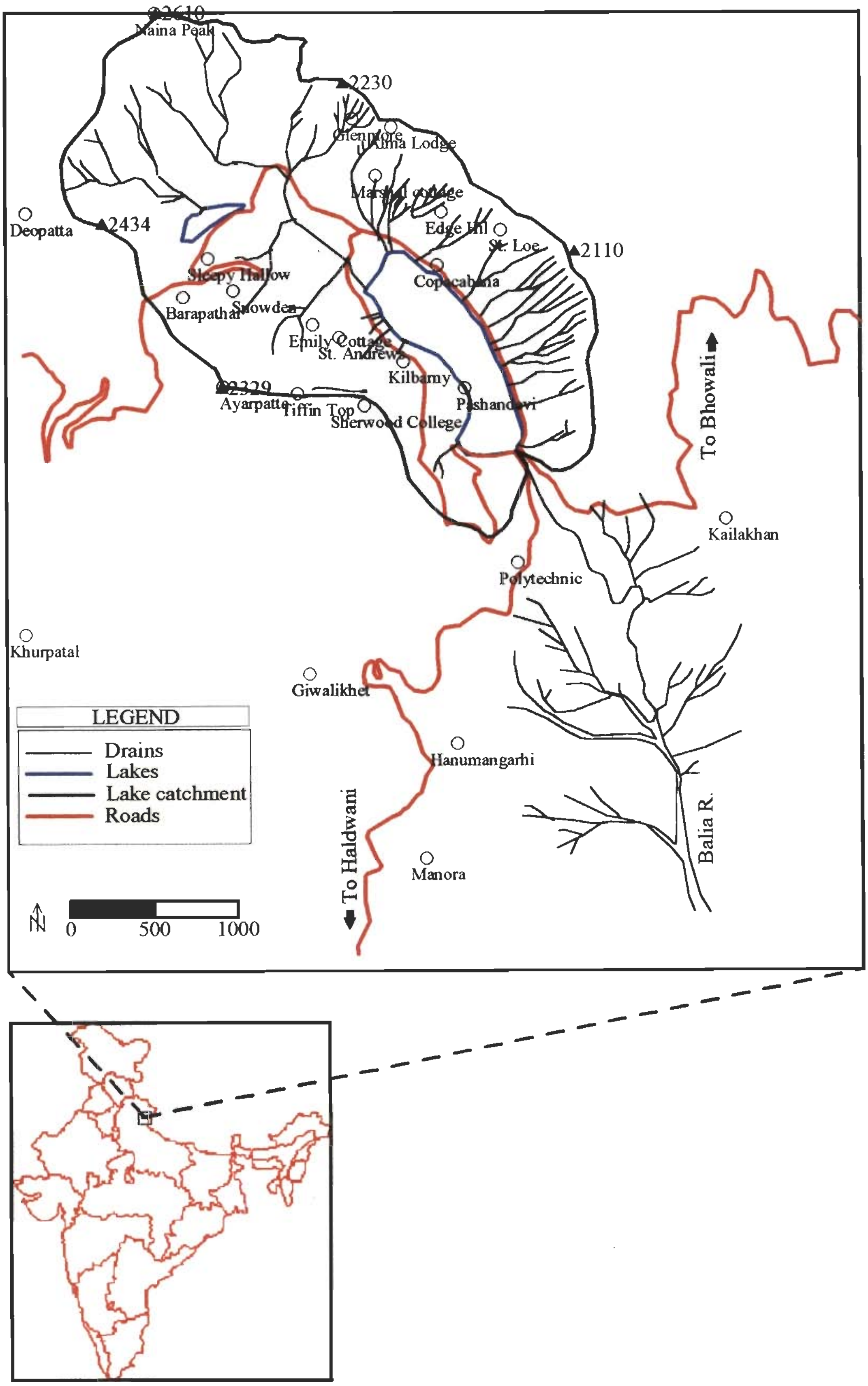


Figure 3.1 Location map of the study area showing Nainital lake

Table 3.1 Lithostratigraphic succession of the Mussoorie Group in Nainital Hills of Krol Belt, Kumaun Lesser Himalaya (152)

Mussoorie Group	Tal Formation	NarainNagar Member	Purple green shale
		Giwalikhet Member	Carbonaceous shale
	Krol Formation	Sherwood Member	Cherty and phosphatic massive dolomite
		Bist College Member	Greywacke
		Pashandevi Member	Dark grey and blue dolomitic limestone.
		Barapatthar Member	Black shales, marlite argillaceous / carbonaceous limestone.
		Hanumangarhi Member	Red and purple shales (pockets of gypsum)
		Manora Member	Calcareous slate, marlite, argillaceous limestone.
	Blaini Formation	Kailakhan Slate	Purple slate, calcareous sandstone, Siliceous - dolomitic limestone.
		Pangot Member	Ash grey / black shale (pyrite + phosphatic nodules)

Krol Formation: Carbonate rocks in and around the Nainital lake have been identified as Krol formation by several workers (101, 68, 9, 10, 11, 74, 107, 108, 56, 149, 150, 151, 152, 131). The Krol formation is divided into six members which are best observed in the Tiffin Top - Baldiakhan section across Ayarpatta ridge. Manora member is well developed on the Manora ridge and can be seen in the cantonment area of the Nainital town - Bhowali region. Hanumangarhi member is well developed in Hanumangarhi temple area, with pockets of gypsum observed in Nihal Valley. Barapathar member is best developed at Barapathar on Kaladungi road near Sariyatal and on the western slope of Hanumangarhi temple. A large part of the northern slope of Ayarpatta including Pashandevi spur on the lake shore is made up of the Pashandevi member, which is the lowest unit of the Upper Krol. Bist college member is a local lenticular facies of the Upper Krol formation around the Ayarpatta area in a semi-circular manner. Sherwood member is characterised by massive dolomite, locally cherty and phosphatic in nature (22) which is well developed in the Ayarpatta ridge and gradually grades into the Giwalikhet member of the Tal formation.

Tal Formation: The Tal formation is divided into two members - lower Giwalikhet member and the upper NarainNagar member.

Dolerite: Dolerite dyke occurs in the west, south-west and south of Ayarpatta peak (108). It is also seen near Alma lodge area between Marshall cottage and Snowview.

3.1.1.2 Structure

The structures of the lake basin and surrounding areas have been studied in detail by several workers (108, 107, 131). The studies show that there are three phases of folding. In the first phase, larger folds trending NW-SE with southwardly overturned asymmetric Ayarpatta - Gairkhet syncline and asymmetric and faulted Deopatta anticline are developed. During the second and third phases, E-W to ENE-WSW folds and NE/NNE - SW/SSW trending gentler folds are developed.

A regional fault, called Lake fault or Naini fault cuts the deformed synclinal massif into two parts (Figure 3.2). The southern part is made of Blaini, Krol and Tal formations and is characterised by southward verging tight folds. The northern part is made of older Jaunsar group and is characterised by overturned folds verging towards north. The steep reverse Naini fault is traceable from Naina and Deopatta peaks through Nainadevi temple along Balia ravine (101).

Parallel to the Naini fault, a sympathetic Ayarpatta fault (101) has considerably deformed the overturned limb of the asymmetrical syncline lying south to it. The bedding plane fault developed between Pashandevi dolomite and Barapathar rhythmite is traceable between St. Andrews and Emily cottage. The whole belt is cut by a large number of N-S and NE/ENE-SW/WSW trending tear faults (152). Two minor faults, the Sleepy Hallow fault and another fault (traceable from Narain Nagar through the Nainital Polytechnic), have affected the Middle Krol Unit leading to the formation of a lakelet behind Snowdon.

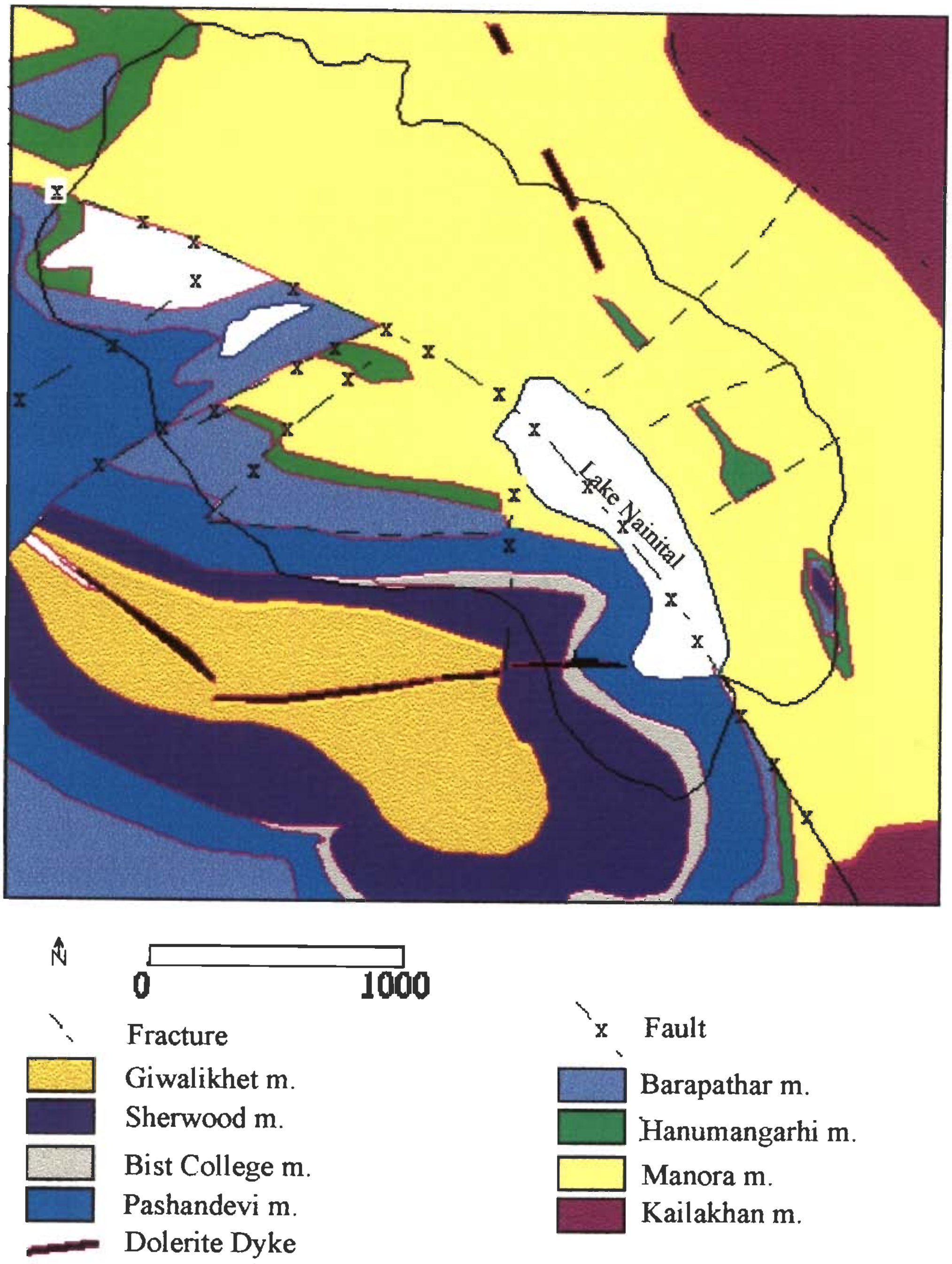


Figure 3.2 Geological settings around Nainital lake (After Valdiya, 1982)

3.1.1.3 Geochemistry

The available literature suggest that the lake basin comprises of shales/slates, dolomites, dolomitic sandstones, purple sandstones, quartzite and dolerite. So far, no detailed work has been done on geochemistry of the rocks of the Nainital basin except for some data based on specific gravity of the rock samples (70). The average specific gravity of the blue-grey dolomite of Nainital basin is 2.82 that corresponds to 64.3% of $CaCO_3$ and 35.7% of $MgCO_3$.

Table 3.2 Chemical composition of dolomitic rocks of Nainital lake basin (70).

Composition	Sandy Dolomite	Dolomitic Sandstone
$CaCO_3$	32.48%	12.98%
$MgCO_3$	28.33%	9.44%
Ferric oxide, Alumina etc.	1.78%	1.88%
Insoluble residue	34.86%	75.76%
Carbonaceous matter	2.95%	-

However, silicates, clay minerals, carbonaceous matter etc. have often been found to be present in considerable quantity. This is evident from the chemical analysis of a sample from Ayarpatta. It shows $CaCO_3$ as 50.13%, $MgCO_3$ as 40.89%, Alumina etc. as 1.63% and insoluble residues as 7.71% (70). The overall chemical composition of sandy dolomite and dolomitic sandstone near East Laggan are given in Table 3.2 (70).

3.1.2 Drainage

The drainage map of Nainital lake basin (Figure 3.1) shows that the area to the west of the lake (Ayarpatta ridge) is marked by very few streams while in the east (Sher-ka-Danda ridge), several parallel to subparallel streams exist. The differences in the fluvial morphometric parameters (Table 3.3) are due to the presence of several lineaments (123). There are around 20 channels that drain the entire Nainital lake basin. Out of these only Nainadevi and Rickshaw stand drains are perennial due to spring discharge and sewage disposal.

Table 3.3 Fluvial morphometric parameters of Nainital lake basin (123)

Geologic formation	Rock type	Morphometric parameters		
		Drainage density (km/km^2)	Stream frequency (Number of channels / km^2)	Bifurcation ratio
Lower Krol	Shale and Marl	6.05	8.15	4.50
Middle Krol	Red Shale and Marl	4.15	5.30	3.20
Upper Krol	Dolomite and Limestone	0.30	1.25	2.00

3.1.3 Land Use Pattern

For land use classification of the lake catchment area, digital SPOT satellite data of November, 1987 and March, 1989 were used. The digital satellite imageries and the base map of Nainital lake catchment area, prepared from the Survey of India toposheet number 53 O/7 (scale 1:50,000) were coupled and processed with ILWIS® and ERDAS® software packages. Minor land use classes such as roads and play grounds have been classified by using the Survey of India Nainital guide map (scale 1:10,000) and forest map of Nainital (scale 1:7875). Ground truth study was carried out to establish the land use. The area under different land uses were measured by using digital planimeter and opisometer. The results are given in Table 3.4. From the land use data, it is seen that the urbanisation and forestry are almost equal and account for about 80% of the total land area.

Table 3.4 Land use data of Nainital lake catchment

Class	Area (km^2)	Percent Area
Reserved forest	0.9932	21.12
Other forests and shrubs	0.9689	20.60
Built-up	1.9195	40.81
Roads	0.1000	2.13
Water bodies	0.4858	10.33
Play ground	0.0527	1.12
Barren land	0.1829	3.89

3.1.4 Soil Properties

To understand the soil moisture characteristics of the area under investigation, soil samples were collected from different parts of the lake basin. Soil moisture retention characteristics were determined using the pressure plate apparatus method (147). The results of the laboratory investigations are presented in Table 3.5. The data show that the average field capacity is around 27% and wilting point around 19%, typical characteristic of a clayey-loam soil.

Table 3.5 Moisture retention characteristics of soil samples collected from different sites in Nainital lake catchment.

Site	Soil Moisture Content	
	Field Capacity (0.15 bars)	Wilting Point (15 bars)
Alma cottage	0.3125	0.1997
Ayarpatta - 1	0.2035	0.0856
Ayarpatta - 2	0.1996	0.1146
Staff House	0.3489	0.2446
Catchment average	0.2661	0.1611

3.1.5 Infiltration

Infiltration process is influenced by several factors such as intensity of rainfall, soil characteristics, soil moisture content, vegetal cover, land use, entrapped air etc. However, one of the major factors is pore size distribution in the soil column. In mountainous terrain, the stream flow is mainly along the valley bottoms. Therefore, the information on infiltration characteristics along drainage channels is vital for understanding the hydrology of the terrain. Hence, infiltration tests were conducted during August, 1998 by using standard ring infiltrometers at three different sites (Figure 3.3) in Nainital lake catchment area. The results are given in Table 3.6.

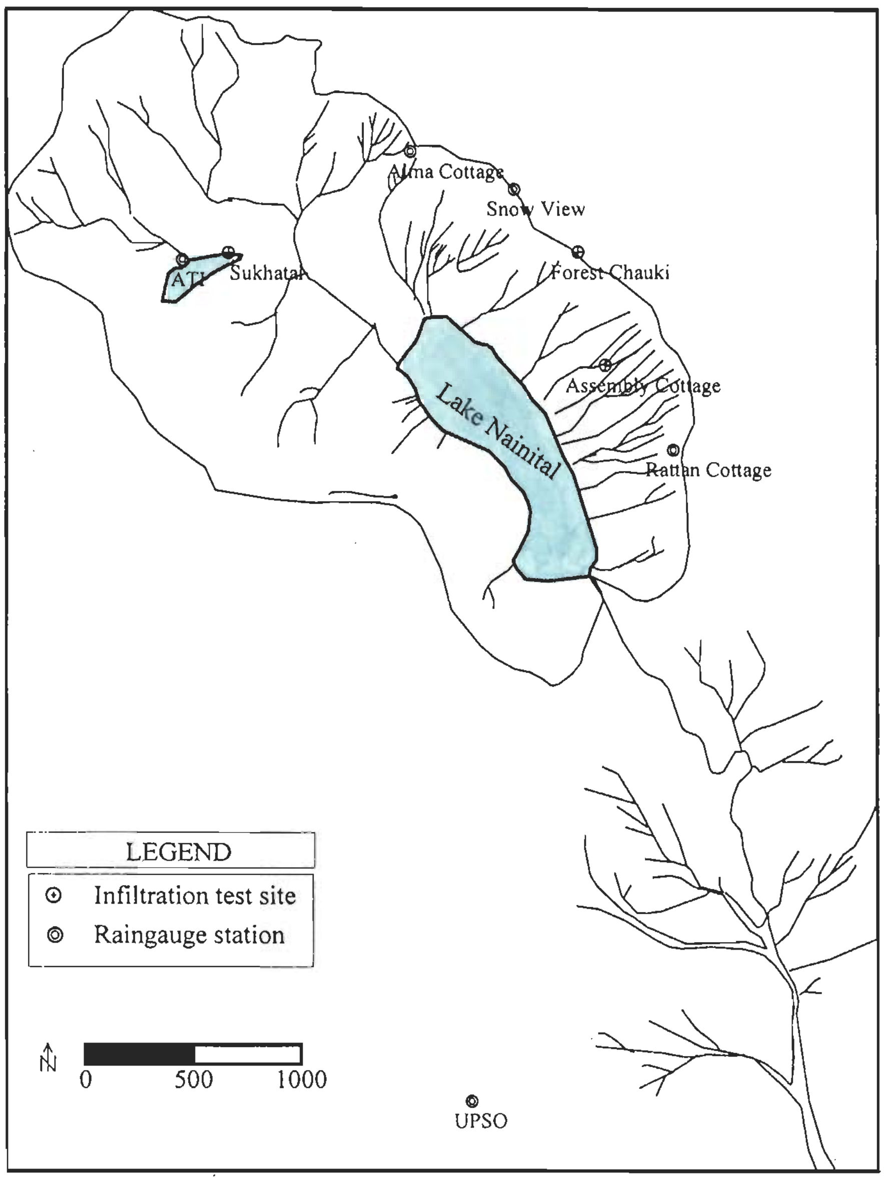


Figure 3.3 Map showing the locations of raingauge stations and infiltration test sites in Nainital lake catchment

Table 3.6 Infiltration rates (f_c) at different sites in lake catchment area

Site	Location	Site description	f_c (cm/h)
Assembly Cottage campus	Near Lakeview Cottage along Birla Vidyamandir Road	Grassy land, slope 15°, macro-pore dominated weathered shale with moderate soil cover	50
Forest Chauki	Along Birla Vidyamandir Road	Thick clayey soil	1.1
Lake bed	Lake Sukhatal	Flat land, macropore dominated, shattered and weathered shale with little soil cover	78

These results indicate that infiltration rate in the lake basin varies widely. The higher infiltration rates along the valley bottom might be responsible for low surface runoff (25-30%) in the Nainital lake catchment. Drying-up of Sukhatal lake at the end of monsoon season may also be attributed to the higher infiltration capacity of Sukhatal lake bed.

3.1.6 Precipitation

In order to understand the precipitation characteristics in the study area, annual rainfall data for Nainital basin were collected from the Uttar Pradesh Public Works Department (UPPWD) for the period 1895-1995, the Uttar Pradesh State Observatory (UPSO) for the period 1965-95 and from the raingauge stations installed by the National Institute of Hydrology at Rattan cottage, ATI and Alma Cottage (Figure 3.3) for the period 1994-95.

3.1.6.1 Long-term average precipitation rates

Hundred and one year annual rainfall data (1895 - 1995) collected from UPPWD indicate that the mean annual rainfall of the basin is 2488 mm, ranging from 1367 mm (1991) to 3910 mm (1910). Time series analysis has been done by autocorrelation method (81), for the Nainital rainfall and the results are plotted in Figure 3.4. The plot of the autocorrelation coefficient, r(k) versus the time lag, k (called a correlogram) shows that no trend is discernible in the annual rainfall series.

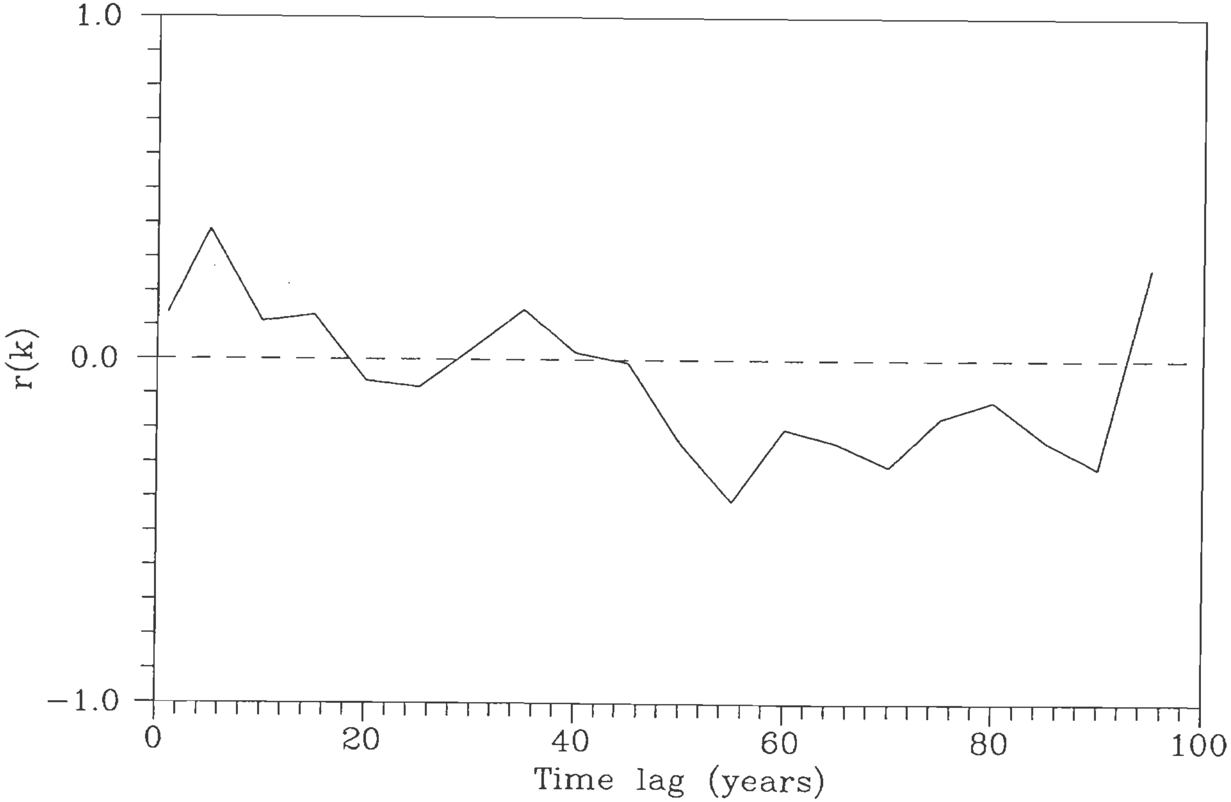


Figure 3.4 Correlogram for annual rainfall of Nainital lake catchment.

3.1.6.2 Seasonal variability

Seasonal variation in precipitation controls the seasonality of stream flow and groundwater recharge. Monsoon regions in particular have pronounced effect of seasonal variation in precipitation. Most parts of the Indian sub-continent are characterised by monsoon precipitation, and the start and end of monsoon season vary from place to place. The monsoon period in the Kumaun Himalaya is generally from mid-June to mid-September. The monsoon precipitation is moderate to heavy and mainly due to moist air-currents from Bay of Bengal. During winter season (January to March), the precipitation is generally light to moderate with occasional snowfall originating from Caspian sea. This type of winter weather system is known as Western - Disturbances.

The average monsoon and non-monsoon rainfall for Manora (Figure 3.5) show that monsoon rainfall plays an important role in the hydrology of Nainital area. Further, it is observed that the monsoon and non-monsoon rainfall contribute around 86% and 14% respectively, to the total annual rainfall.

3.1.6.3 Orographic effects on precipitation

Non-uniform precipitation in mountainous regions is attributed to the interaction of weather systems with topography. This is known as orographic effect on precipitation. It has been observed that there is remarkable variation in rainfall with altitude in Outer, Middle and Greater Himalaya (134). In the Outer Himalaya, during the monsoon season, the rainfall in windward side increases upto an altitude of 600 m and then decreases while during the other seasons, the amount of rainfall increases linearly with altitude. These are attributed to the difference in weather conditions during monsoon and non-monsoon seasons. In the present study, monthly rainfall data of three stations viz. Manora, ATI and Snowview have been used to infer the orographic effect on precipitation. From the data it appears that with an increase in altitude, the rainfall decreases during July and August, but increases in September. This observation is substantiated by variation in the 30-year (1966-95) mean monthly relative humidity observed at Manora (Figure 3.6). The mean monthly relative humidity is higher during the months of July and August and relatively lower in September. However, due to smaller altitudinal range in the lake basin (less than 600 m) the orographic effect is not discernible.

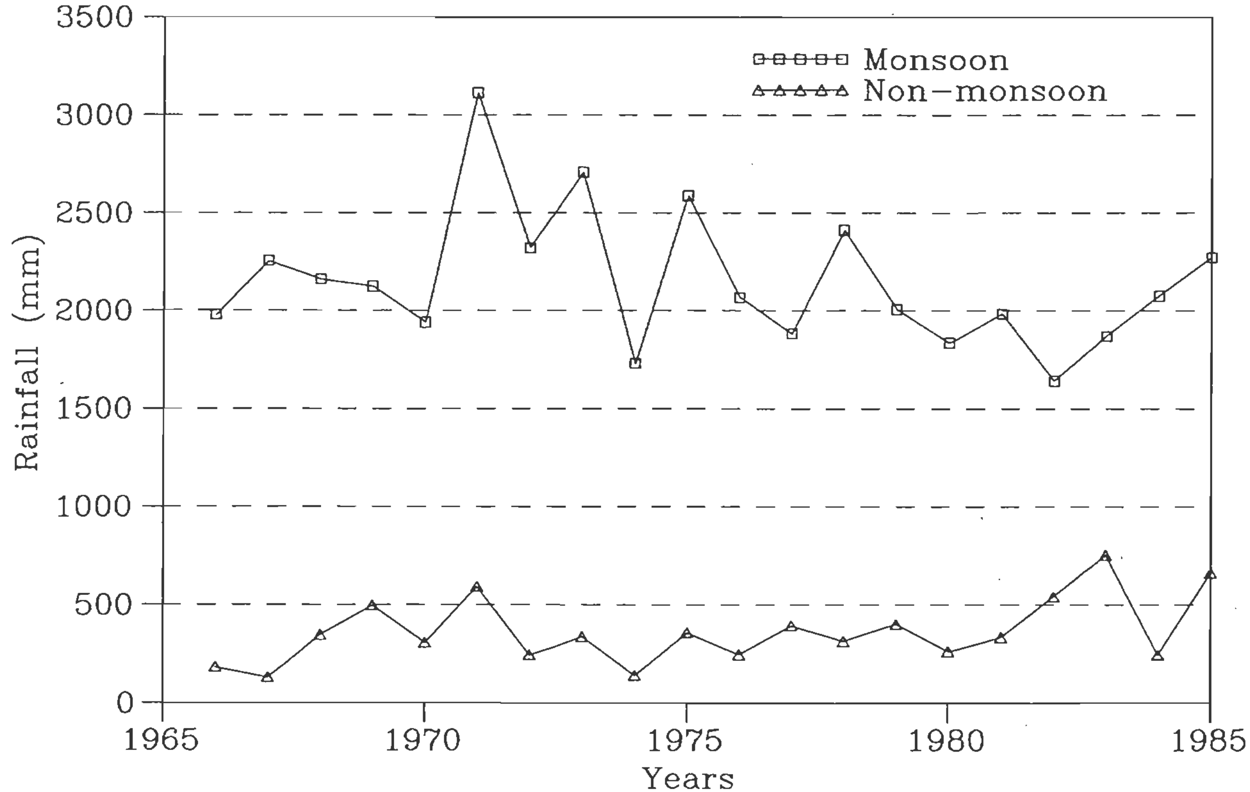


Figure 3.5 Variation of monsoon and non-monsoon rainfall at Manora (Nainital).

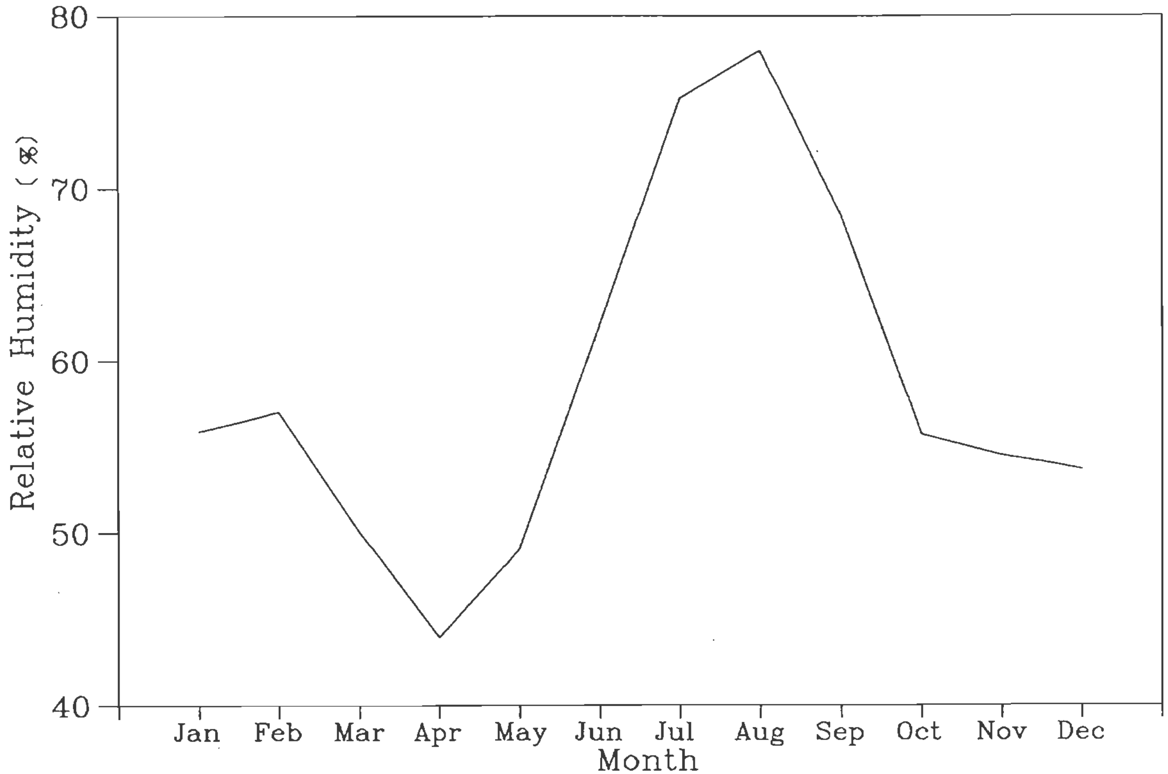


Figure 3.6 Variation of 30-year average monthly mean relative humidity at Manora (Nainital)

3.1.7 Hydrogeology

Systematic hydrogeological studies in Nainital basin have never been conducted earlier, mainly due to the abundant surface water availability. However, investigations conducted on landslides and land hazard zonation by few investigators (73, 8) have provided some insight in understanding the probable hydrogeological setting of the basin. As mentioned earlier, most part of the basin is characterised by calcareous and pyritiferous shales. Most of the lake catchment area lying north of Lake fault, is characterised by Manora shales that dip towards the lake. Average dip of the shales/slates in Sher-ka-Danda is 40° WSW and in Alma Hill 40-48° SW. The main causes of slope instability are rain water seepage through bedding planes and rock formations in eastern and western part dipping towards the lake (70).

Infiltrating rain water in the Nainital lake catchment seems to be moving preferentially along bedding planes. It appears that infiltrating water does not percolate deeper, but discharges either into the lake or run as seasonal springs. During the present investigations, it has been found that several springs are seasonal. Their discharge diminished with time and they completely dried by the middle of post-monsoon season.

To understand the groundwater condition in the basin, a cross-section based on the available lithologs from 4 bore-holes drilled by UP Department of Mining and Geology (8), has been prepared. The ENE-WSW cross-section is drawn between St. Loe, in the Sher-Ka-Danda Hill and the Copacabana Restaurant near the lake (Figure 3.7). The top few meters are characterised by overburden, which is underlain by weathered shales of Lower Krol upto a depth of 35 m with an intervening shear zone between 29 and 34 m. The depth of the overburden increases towards the WSW part of the section i.e. near the foot-hills. The northern part of the lake is characterised by landslide debris, nala deposits and lake deposits upto a depth of 116 m. Below this depth, shales/slates of Middle Krol have been reported (8).

It is evident from above discussion, that Nainital lake basin is structurally and stratigraphically a complex area, but essentially consists of carbonate and shale formations. Shales are the least productive rocks and carbonate rocks vary widely in their water yielding capacity. This is due to solvent action of circulating water, particularly in folded limestone, the water may move in the down dip direction from the recharge area to the discharge area. The

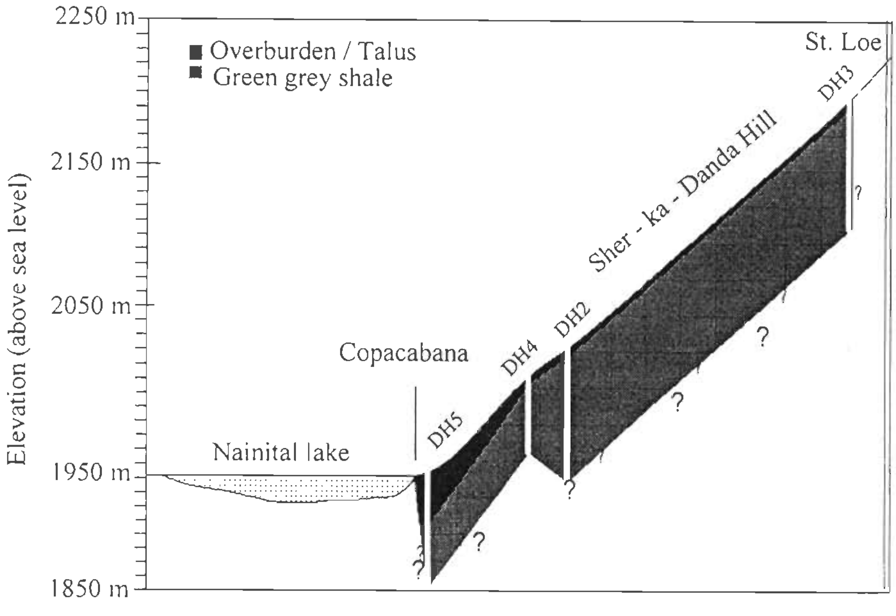


Figure 3.7 Geological cross-section along Copacabana and St. Loe in the Sher - ka - Danda hill, Nainital. Data source: Ashraf, 1978

joints and bedding planes have a pronounced effect on the solution patterns and movement of water in carbonate rocks. In carbonate terrain, faults affect the lateral movement of water especially if water bearing limestone beds are faulted against relatively impervious rocks (141). This condition leads to development of springs along the fault zone. The perennial Pardhadhara spring located in the Nainital basin, is probably a result of Naini fault that has brought the limestone of Middle Krol against the Manora shales of Lower Krol. Similar to this, there may be several sub-surface springs emerging in Nainital lake along Naini fault.

The dolomite occurring in lake catchment is marked by swallow holes, such as the one found north of Sherwood College. There are a few cave-like structures observed in and around the lake basin near Pashandevi temple, Gupha Mahadev Temple and in the northern slope of Sukhatal. However, massive dolomite and associated limestones have not developed all the characteristic features of karst topography.

3.1.7.1 Aquifer properties

Aquifer tests in Manora shales of Sher-ka-Danda slope near Spring Cottage (above the spring) have been conducted by slug test method. The slug test was performed in a small diameter monitoring well using Hvorslev method (49). The method involves addition of a known quantity of water (slug) into the well casing. As a result of the introduction of slug the water level rises and after certain time, it gradually falls back towards the static water level. The water levels are measured in the well, before and immediately after the introduction of slug and also at fixed intervals till the water level falls back to the static level. The height to which the water level rises above the static water level immediately upon introduction of the slug is h_0. The height of the water level above the static water level at time t, after the introduction of slug is h. If the ratio h/h_0 is plotted against the corresponding time (t) on a semi-logarithmic paper, the time - h/h_0 data falls on a straight line. If length of the piezometer is more than 8 times the radius of well screen, then the hydraulic conductivity (K) in m/day is determined by the following equation (49):

$$K = \frac{r^2 \ln (L_e/R)}{2L_e T_0} \tag{3.1}$$

where, r = radius of the well casing (m)

R = radius of the well screen (m)

L_e = length of the well screen (m) and

T_0 = time taken for water level to fall to 37 percent of the initial change (day).

In the present study, an existing piezometer (DH#2) has been used. The total length of the well is 91.72 m and the radius is 0.019 m. The static water level in the well prior to the test was 17.00 m below ground level. A known quantity of slug (15 L of water) was introduced into the well. The water level measurements were taken using a quick-response water level indicator with an accuracy of ±1.0 mm. The water level rose to a height of 0.29 m immediately after the introduction of slug. The total time taken for the water level to fall back to the static water level was around 9.75 minutes and T_0, calculated from the h/h_0 - t semi-log plot (Figure 3.8), was 5.18 minutes or 0.003597 days. The hydraulic conductivity has been estimated by using Equation (3.1) as 4.64 * 10^{-3} m/day or 5.4 * 10^{-8} m/s. Assuming an aquifer thickness of 100 m, the aquifer transmissivity (T) has been estimated as 0.464 m^2/day. The estimated K and T values for Manora shales are relatively less than the shale formations occurring in other parts of India (78). However, the estimated K value is slightly higher than the range of values (2 * 10^{-9} m/s to 1 * 10^{-13} m/s) quoted for shales (42).

The storage coefficient of the aquifer is estimated from the well hydrograph analyses. Generally, the well hydrograph follows a trend similar to that of a stream hydrograph, with the rising and recession limbs corresponding to periods of recharge to and discharge from groundwater storage. The recession limb is characterised by an initial steep slope representing the rapid drainage from storage, followed by a gentle limb that tends to stabilise after a prolonged period of drainage. The physical process of releasing water from storage in the aquifer may be defined as (78):

$$(h - h_m) = (h_0 - h_m)\, e^{-\alpha t} \tag{3.2}$$

where, h = water level in the well at time t

h_0 = water level at the start of the recession

h_m = water level when the rate of recession is nil (i.e. at the end of post monsoon period)

α = recession coefficient, may be approximated to S (storage coefficient).

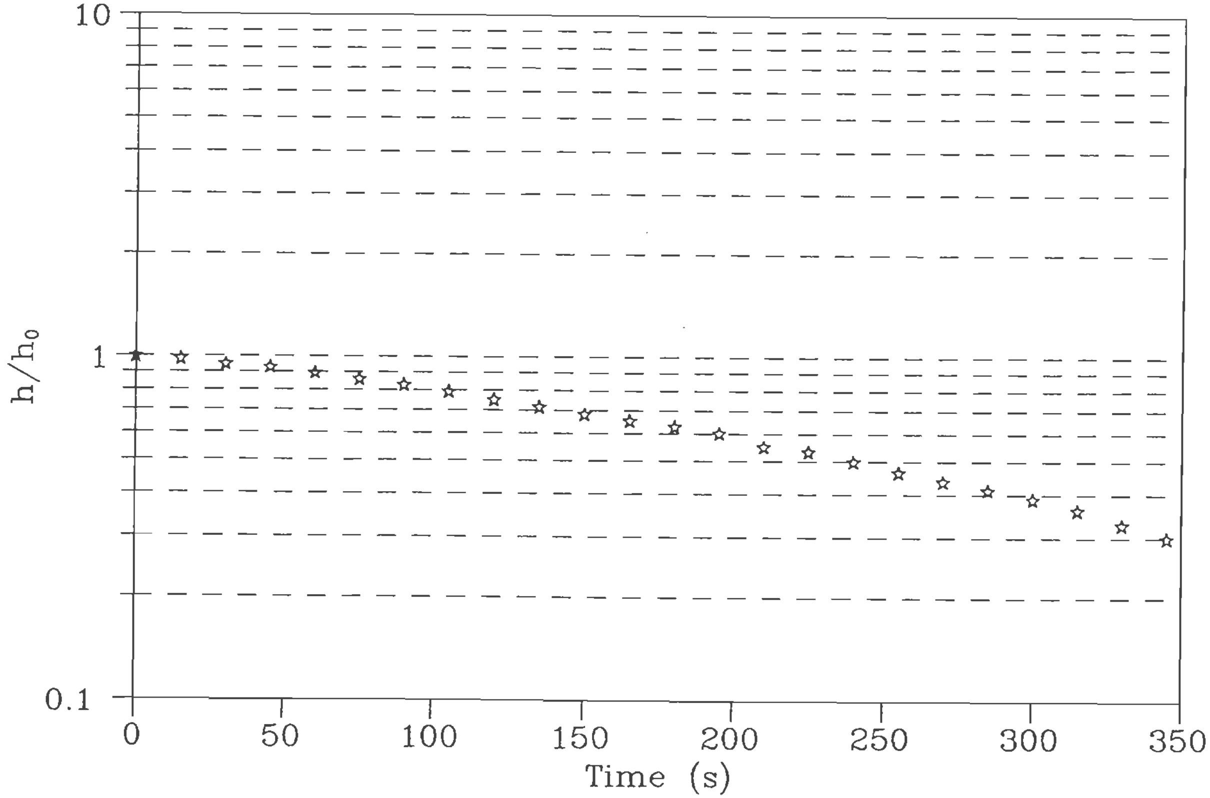


Figure 3.8 Plot of h/h_0 versus time for the estimation of K in Manora shale by slug–test method.

If the values of h and h_0 are taken with reference to h_m, the above Equation may be re-written as:

$$h = h_0 e^{-\alpha t} \tag{3.3}$$

Rearranging the above equation, we get

$$\alpha = \frac{\log (h_0) - \log (h)}{t} \tag{3.4}$$

From this the value of α and therefore, S of the aquifer may be estimated. The groundwater level data pertaining to the period 1990-91 for wells DH#2 and DH#4 were obtained from UPPWD and analysed using Equation 3.4. The estimated S ($\approx \alpha$) are found to be $1.56 * 10^{-4}$ and $1.34 * 10^{-4}$ for wells DH#2 and DH#4 (well hydrographs are presented in figures 3.9a and 3.9b). Since the estimated values are very close to each other, a mean value of $1.45 * 10^{-4}$ may be considered as true representation of the aquifer storage coefficient. The specific yield of Manora shales (around 0.015%) is much lower as compared to other shale formations [Cuddapah shales, Andhra Pradesh have a mean specific yield of 3%, (78)]. The results indicate that Manora shale has very low productivity and therefore, groundwater movement may occur only through lineaments occurring in the shale.

3.1.7.2 Groundwater withdrawal

Domestic water supply to the Nainital town is met through pumping of groundwater from deep tube wells and an open well located at northern banks of Nainital lake and from few natural springs. Pumping stations are operated and the records are maintained by UP Jal Nigam, Jal Sansthan, and Army authorities at Nainital. Discharge capacities of the pumps installed at lake banks are given in Table 3.7.

The total monthly draft through pumping has been calculated for the period from January 1994 to December 1995 and is given in Table 3.8. The monthly discharge varied from 2,23,580 m^3/month to 4,99,100 m^3/month during 1994 and 1995. It is observed that pumping was higher in April to June and lower in September to November.

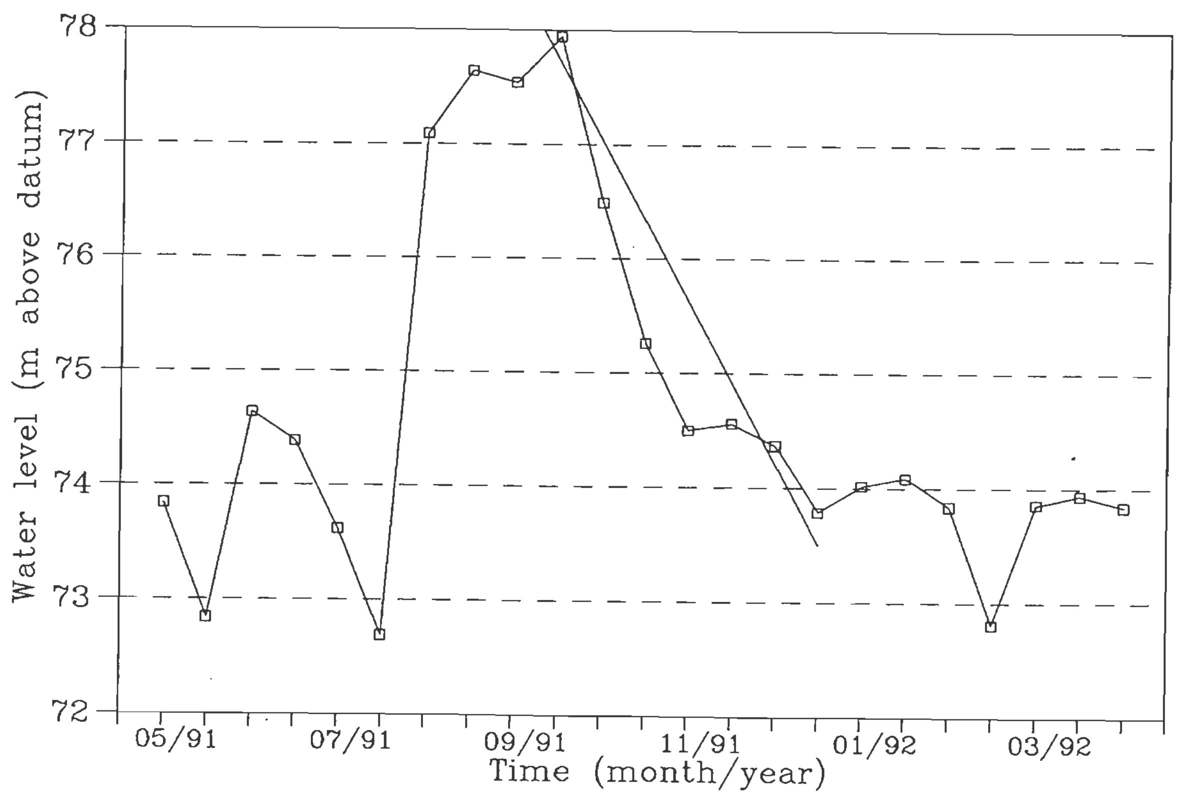


Figure 3.9a Well hydrograph of DH#2 located in Manora shales below the spring, near Spring cottage, Nainital

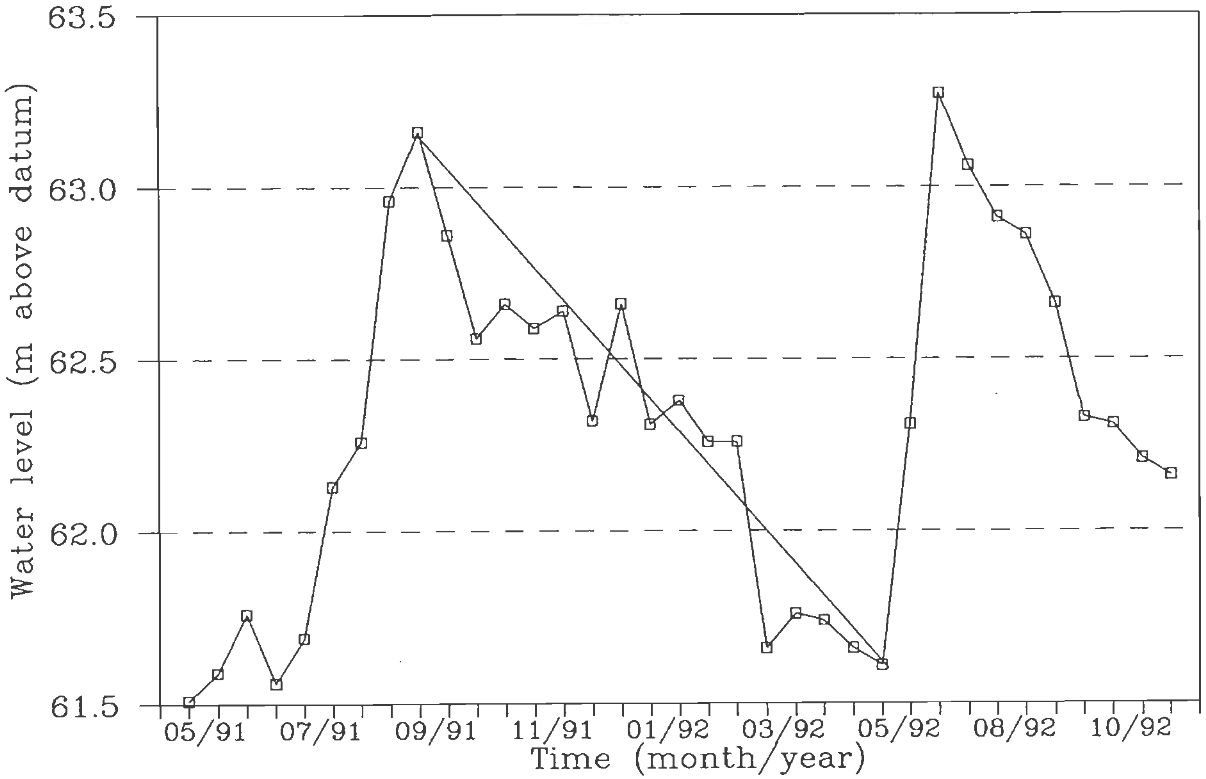


Figure 3.9b Well hydrograph of DH#4 located in Manora shales above the spring, near Spring cottage, Nainital.

Table 3.7 Discharge capacities of wells located at Nainital lake banks. (Data collected from UP Jal Nigam, UP Jal Sansthan, and Army authorities).

Pump No.	Location	Year of installation	Pump Capacity (m^3/h)
1	Open well	1923	69.00
2	Open well	1989	120.00
3	Open well	1985	142.20
4	Open well	1958	95.40
5	Open well	1958	72.00
6	Open well	1958	72.00
7	Open well	1958	95.40
8	Open well	1983	244.80
TW1	Tube well-1; Jal Sansthan	1991	90.00
TW2	Tube well-2; Jal Sansthan	1991	132.00
TW3	Tube well-3; Jal Nigam	1994	204.00
MW1	Army Pump House	1980	360.00

Table 3.8 Quantity of water pumped (m^3) from the banks of Nainital lake in different months during the years 1994 and 1995.

Months	1994	1995	Months	1994	1995
January	244470	372970	July	402760	440690
February	243550	356250	August	268120	321450
March	295500	409180	September	223580	242750
April	334900	479270	October	318820	234330
May	459290	499100	November	334170	295700
June	469130	468240	December	340930	331990

Since the tube wells and open well are located on the banks of the lake (5 to 10 m from lake water) characterised by highly permeable land slide debris, lake and nala deposits, the groundwater is replenished by lateral seepage from the lake. Therefore, the pumped water is a mixture of groundwater and lake water, which may be assessed from the isotopic characteristics of the pumped water.

3.1.7.3 Springs

Eleven springs are found associated with major fractures of Manora shales from Sher-ka-Danda - Snowview - Alma Hill slopes on upstream side of the lake. The discharge data of these springs for the period 1976-79 were collected from UPPWD. Using the historical discharge data, the springs have been classified into two groups (Table 3.9).

The classification of the springs has been done on the basis of Coefficient of Variation (CV) of monthly discharge. Therefore, based on monthly discharge rates, the springs are classified into Group I (CV < 70%) and Group II (CV > 70%). Group I has a fairly larger mean monthly discharge (> 20 l/min.) than those of Group II (< 5 l/min.), with an exception of Alma Spring. The difference in the discharge rates could be due to the differences in the volume of corresponding groundwater reservoirs.

Table 3.9 Classification of upstream springs based on average discharge and coefficient of variation in the discharge. (Data collected from UPPWD)

Group No.	Spring	Elevation (m above m.s.l.)	Mean Discharge (l/min)	Coefficient of Variation (%)
Group I	Spring Cottage	2015	25	36
	Lakeview	2040	21	56
	Drain #14	2075	20	62
Group II	Oak Lodge	2050	4	73
	Drain #3	-	5	77
	Alma Spring	2130	21	78
	Kumaun Lodge	2060	4	82
	Doctor House	1960	4	83
	Chunadhara	2035	4	85

In terms of discharge the largest spring within the lake catchment is the Pardhadhara, (the word dhara denotes a spring) located in the vicinity of Lake fault. It is considered as a karst spring from dolomitic country rock, later cut by the Lake fault (153). The CV, mean and maximum discharge of Pardhadhara are 53%, 1540 l/min. and 2613 l/min. respectively. The larger discharge rate of Pardhadhara may be due to replenishment from the Sukhatal catchment area and the adjacent limestone terrain.

In the downstream side of Nainital lake, there are around 15 springs. Amongst these springs, only the Balia ravine springs are of significance due to their larger discharge. The discharge data for some of the Balia ravine springs from earlier records (1948-52) supplied by UPPWD and those monitored during the present study (1994-95) are given in Table 3.10.

Table 3.10 Discharge data of Balia ravine springs.

Spring	Elevation (m above m.s.l.)	Discharge (m^3/day)		CV (%)	Remarks
		Mean	Max		
Dhobighat	1850	160	273	25	Data pertaining to the period 1948-52.
Patwadanger	1830	92	194	48	
Tallital	1900	134	293	49	
Rais Hotel	1750	3385	8056	54	
Sipahidhara	1825	4920	12953	65	
Sipahidhara	1825	656	1660	73	Data collected during 1994-95.

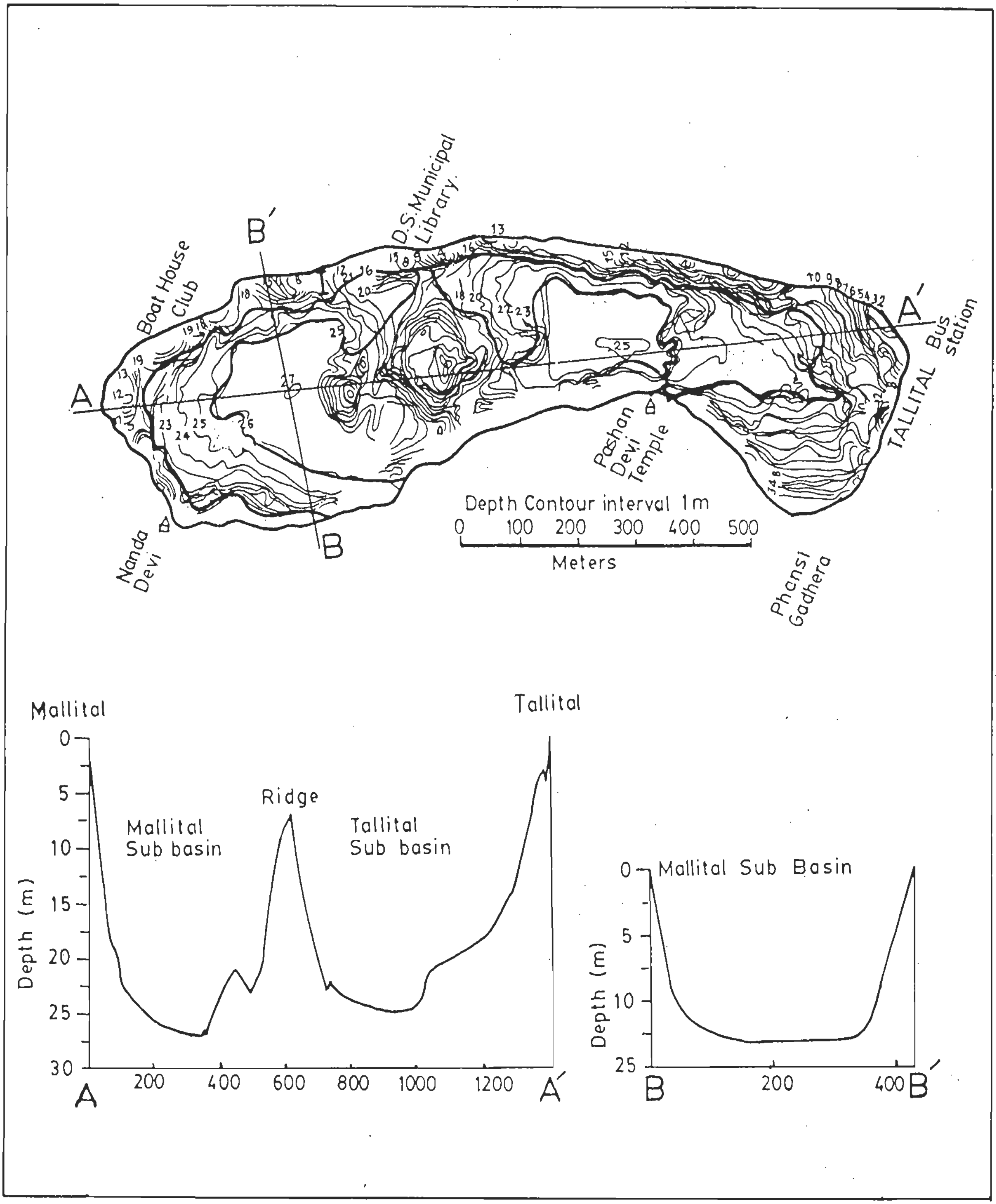
It is noticeable that the discharge of Sipahidhara spring has been reduced considerably between 1948-52 and 1995. Likewise, many such springs in the area show a reduced discharge or have become dry, e.g. Fairy Hall spring. The probable reason for reduction in the discharge of the springs is sedimentation in the Nainital lake, and the consequent clogging of sub-terranean pathways (89).

3.1.8 Characteristics of Nainital Lake

3.1.8.1 Morphology of the lake

The lake is bound, in the east by the Sher-ka-Danda hill, in the north by the landslide deposit called Flat, in the west by the Ayarpatta hill and in the south by Balia ravine. The western and northern banks of the lake are steep, while the eastern bank is slightly shallow due to deposition of sediments by drains. The largest deltaic type sedimentation is found near Nainadevi Temple.

Bathymetric surveys of Nainital lake have been carried out by several investigators using conventional sounding and echo-sounding methods (70, 122, 66). The detailed bathymetric map of Nainital lake floor (66) is presented in Figure 3.10. The lake is divided into two sub-basins



3.10 Bathymetric map of Nainital lake (After Hashimi et al., 1993).

viz. northern Tallital and southern Mallital sub-basins by an 100 m wide underwater ridge at depths ranging from 7 to 18 m below surface. The two sub-basins are roughly of the same size, and the maximum depth in the northern Mallital and southern Tallital sub-basins are 27.3 m and 25.5 m respectively. Further, it is seen from Figure 3.10 that the dolerite trap found in the south-western side of the lake catchment extends upto Lake fault.

3.1.8.2 Estimation of lake volume

Lake volume (V) can be computed by using the capacity - elevation curve or alternatively by using the prismoidal formula (168). In the latter method, the following relation is used:

$$V_{A_1 A_2} = \frac{h}{3} \cdot (A_1 + A_2 + \sqrt{A_1 A_2}) \qquad (3.5)$$

where, h is the contour interval, A_1 is the area enclosed by the upper depth contour and A_2 is that enclosed by the lower depth contour. Summation of results of repeated successive applications of the above equation will yield the total lake volume.

In the present work, volume of the lake has been calculated from the bathymetric contour map (66). The measurement of area of each contour (from 4 m to 27.3 m depth) was carried out by using digital planimeter. Since the contours for shallower depth zone (upto 4 m) were not readable from bathymetric map, the data from UPPWD (19) have been used in the calculation. The calculated area and volume corresponding to depth of the lake are presented in Table 3.11 and the variation of lake area and volume with respect to depth are shown in Figures 3.11 and 3.12 respectively.

3.1.8.3 Thermal regime of the lake

Thermal regime of the lake is a result of heat and momentum transfer across the surface of the lake and gravitational forces acting within the lake due to density differences. The temperature difference between the top and bottom of the lake gives rise to mixing currents, which are limited to small depth cycle in summer, but to greater depths in winter. Thus, the lakes of medium depth remain well mixed during winters and stratified during summers. This phenomenon is a consequence of several meteorological and other factors.

Table 3.11 Volume and total surface area of Nainital lake at different depths

Depth (m)	Cumulative Volume (m^3)			Total lake surface area (m^2)
	Tallital	Mallital	Total lake	
0	-	-	8581714	463365
0.5	_	_	8336862	459730
1.5	_	_	7880420	451721
2.5	_	_	7434634	441625
3.5	_	_	6994371	439340
4.5	_	_	6556175	437052
5.5	_	_	6120267	434765
6.5	_	_	5686646	432478
7.5	_	_	5255311	430192
8.5	_	_	4828296	423846
9.5	_	_	4409556	413653
10.5	_	_	4001588	402307
11.5	_	_	3608547	383846
12.5	_	_	3233679	365961
13.5	_	_	2876212	349038
14.5	_	_	2534507	334423
15.5	_	_	2206547	321538
16.5	_	_	1893896	303846
17.5	_	_	1596355	291280
18.5	418484	897493	1315977	269615
19.5	311653	746553	1058206	245000
20.5	224215	605239	829454	211346
21.5	154637	474653	629289	186376
22.5	96174	352889	449063	169230
23.5	47413	240246	287659	147115
24.5	12371	144294	156665	112307
25.5	-	69333	69333	67307
26.5	-	15846	15846	40769
27.3	-	133	133	769

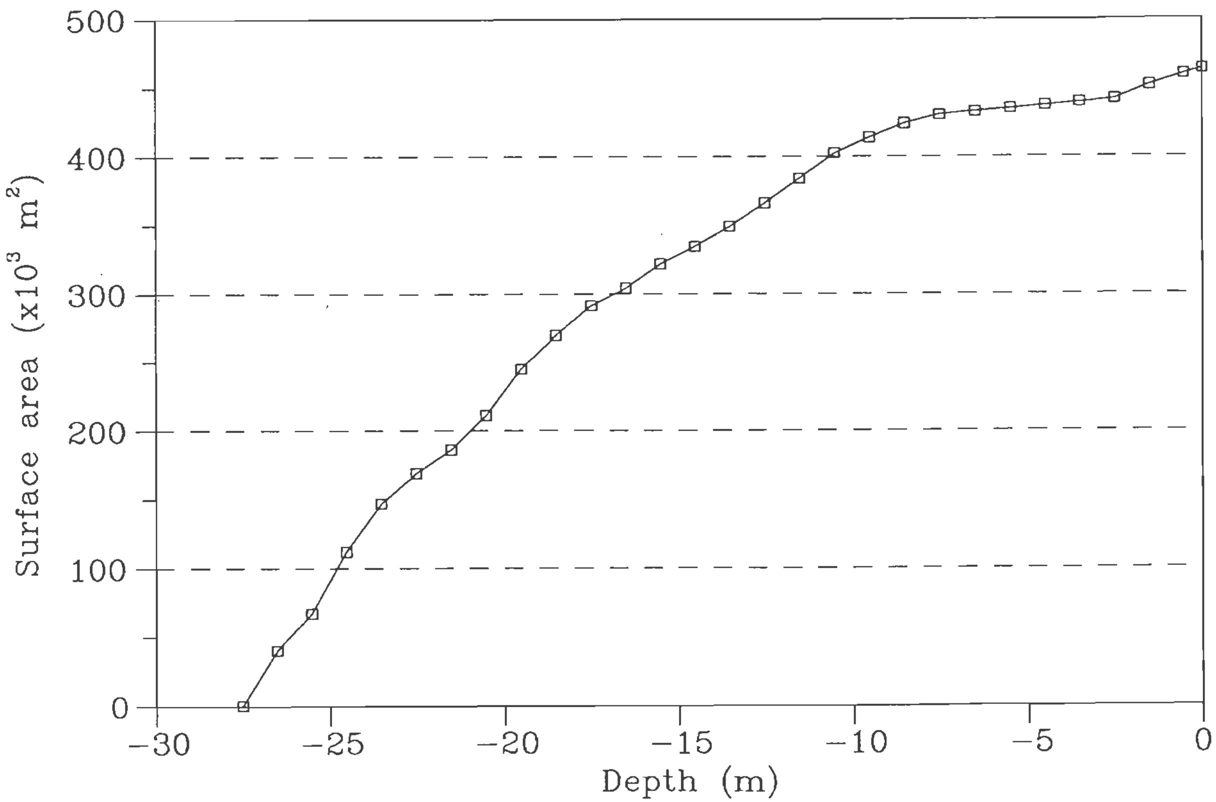


Figure 3.11 Variation of surface area of Nainital lake with depth.

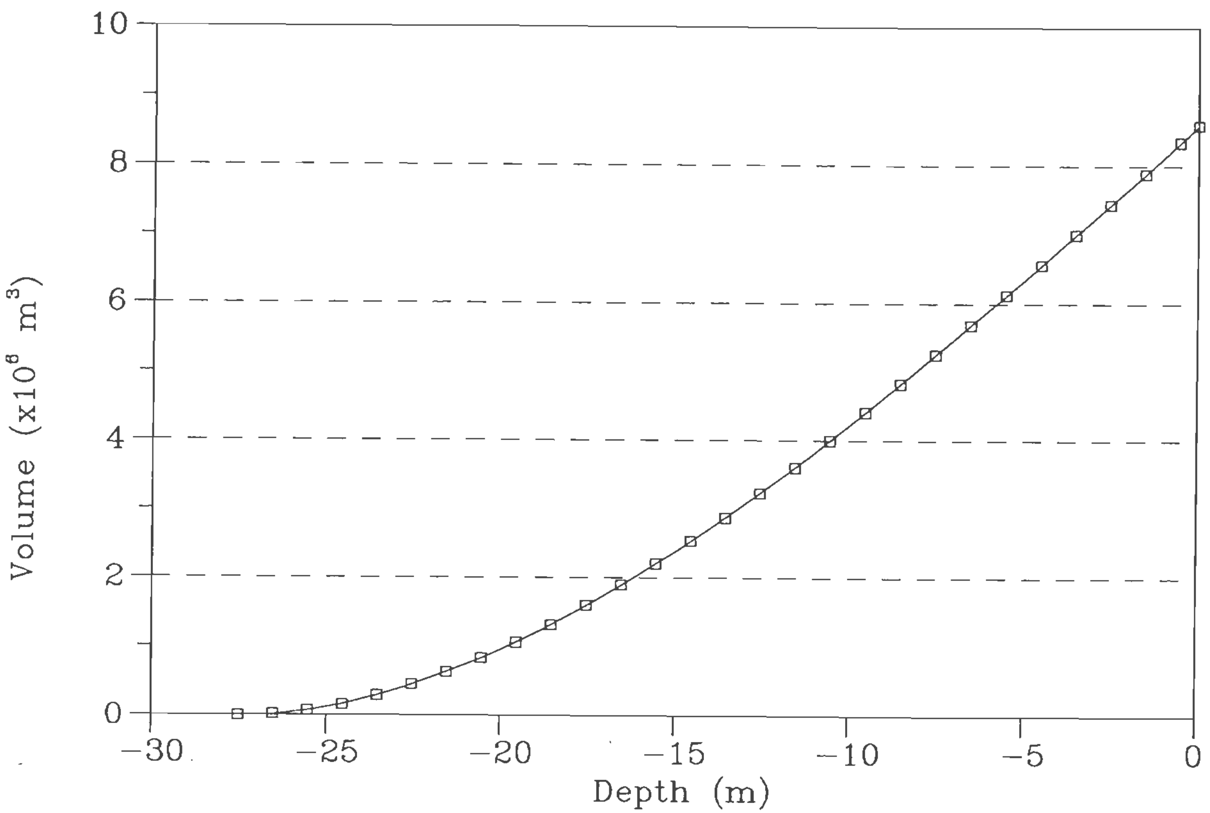


Figure 3.12 Variation of volume of Nainital lake with depth.

During winters, the temperature difference between surface and bottom water is negligible and the lake is considered as well-mixed. During springs, as a result of rise in air temperature, surface water temperature increases producing thermal stratification. This is intensified in summers due to increase in solar radiation. Mixing due to wind action is confined only to the surface layer (called epilimnion) and does not extend upto the bottom layer (called hypolimnion) where the lake becomes stably stratified. The layer, that separates epilimnion and hypolimnion layers, is called mesolimnion. The plane of higher temperature gradient within the lake is called as thermocline. The stable stratification that persists during monsoons, tends to become unstable in early winters, mainly because of falling air temperature thereby resulting in net heat loss. During this period, strong vertical mixing (convection) takes place. The convection and wind action lead to complete mixing of lake called "winter overturn".

Lake temperatures at different depths were measured in both sub-basins of the lake during the period 1994 and 1995. Measurements were made at 3 m depth intervals using an in-built thermometer in the water sampler. The accuracy of the measurement is ±0.2°C. The mean temperature of the lake at different depths, during different months are given in Table 3.12. The temperature variation with depth in Mallital and Tallital sub-basins are plotted in Figures 3.13a and 3.13b. The epilimnion temperatures range from 9.2°C to 22.4°C (Tallital) and 8.4°C to 23.6°C (Mallital), whereas the hypolimnion temperatures range from 8°C to 9.4°C (Tallital) and 8°C to 9.2°C (Mallital). The data indicates that the lake is thermally well-mixed during winters (December to February). The thermocline is developed during March and the lake becomes stratified in April. It gets intensified during May with a maximum difference in temperature between epilimnion and hypolimnion layers observed during June. The stratification continues upto November and the thermocline disappears during the third week of December, leading to winter overturn. The periods of mixing and stratification of the lake are also evident from the vertical exchange coefficients calculated for the lake.

The vertical exchange coefficients across the thermocline has been estimated using the following equation (124). The results are given in Table 3.13.

Table 3.12 Temperatures (°C) observed in the two sub-basins of Nainital lake during 1994-95

Mallital (Northern) Sub-basin									
Depth (m)	0.0	3.0	6.0	9.0	12.0	15.0	18.0	21.0	24.0
January	9.4	9.2	9.2	9.2	9.2	9.2	9.2	9.2	9.2
February	8.4	8.2	8.1	8.1	8.0	8.0	8.0	8.0	8.0
March	15.4	12.4	9.6	8.6	8.3	8.2	8.2	8.2	8.2
April	18.8	17.8	12.4	9.8	9.2	9.0	9.0	9.0	9.0
May	18.6	18.2	15.8	10.2	10.1	9.8	9.8	9.2	
June	23.6	21.2	14.0	9.5	8.8	8.8	8.8	8.6	8.6
July	22.2	22.0	18.0	10.4	9.8	9.8	9.4	9.2	9.2
August	22.2	20.2	18.4	11.0	9.2	9.0	8.8	8.8	8.4
September	20.2	19.3	20.0	11.7	10.0	9.4	9.4	9.0	8.9
October	20.5	20.2	18.1	12.9	10.9	10.0	10.0	9.8	7.5
November	13.6	13.6	13.4	13.2	9.8	9.3	9.3	9.2	
December	11.6	11.0	11.0	11.0	10.8	10.4	10.2	9.0	9.8
Tallital (Southern) Sub-basin									
Depth (m)	0.0	3.0	6.0	9.0	12.0	15.0	18.0	21.0	>21.0
January	9.6	9.4	9.4	9.2	9.2	9.2	9.2	9.2	9.2
February	9.2	8.2	8.2	8.2	8.2	8.1	8.1	8.0	8.0
March	16.0	12.2	9.4	8.6	8.2	8.2	8.2	8.2	8.2
April	19.0	18.0	12.6	9.8	9.2	9.2	9.0	9.0	9.0
May	19.9	18.9	14.8	10.4	9.8	9.4	9.4	9.4	
June	22.4	21.4	14.4	9.4	9.0	8.8	8.7	8.6	
July	22.0	22.0	17.8	10.0	10.0	9.8	9.6	9.6	9.4
August	21.6	20.0	18.4	11.6	9.4	9.2	9.0	8.8	8.8
September	20.2	18.8	17.8	12.4	9.7	9.2	9.2	9.0	9.0
October	20.1	19.6	18.6	13.5	10.8	10.0	9.9	10.2	10.0
November	14.2	13.2	13.3	13.2	10.2	9.4	9.4	9.3	9.2
December	11.4	11.2	11.0	10.8	10.8	10.4	10.2	10.2	10.0

Figure 3.13a Temperature variation in Mallital sub-basin of Nainital lake during 1994-95

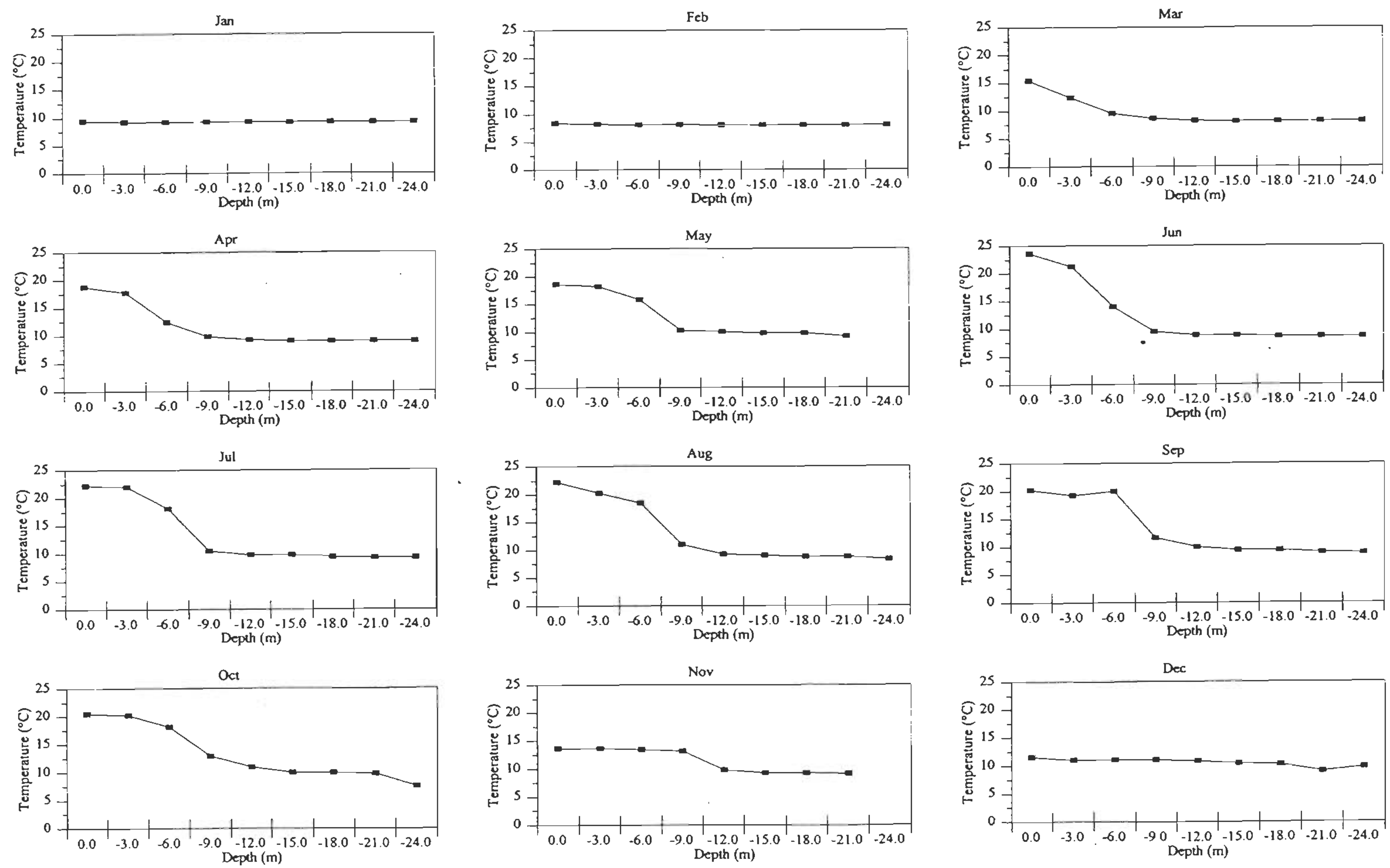
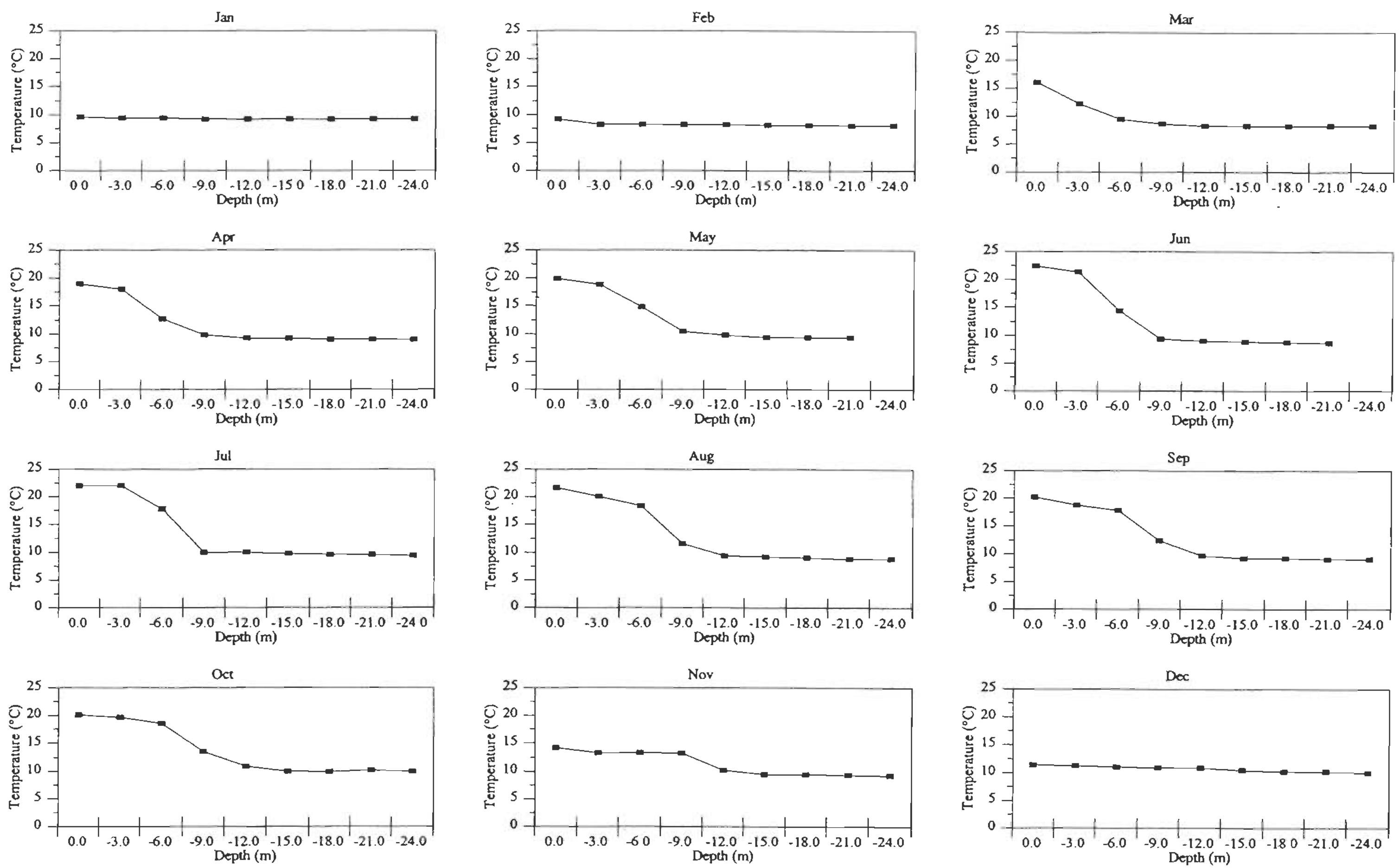


Figure 3.13b Temperature variation in Tallital sub-basin of Nainital lake during 1994-95



$$V_t = \frac{V_h}{2A_t\,\Delta_t}\left(\frac{T_h^{n+1} - T_h^{n-1}}{T_e^n - T_h^n}\right) \qquad (3.6)$$

where, V_t = vertical exchange coefficient (m/d)

V_h = volume of hypolimnion (m^3)

A_t = area of thermocline (m^2)

Δ_t = time interval between temperature measurements (d)

T_h^{n+1}, T_h^n, T_h^{n-1} = temperature of hypolimnion at times n+1, n, n-1 (°C)

T_e^n = temperature of epilimnion at time n (°C).

Table 3.13 Calculated vertical exchange coefficients for the Nainital lake

Month	V_t (m/d)	Remarks	Month	V_t (m/d)	Remarks
January	-	well-mixed	July	0.003	Stratified
February	-	well-mixed	August	0.007	Stratified
March	0.017	Weakly Stratified	September	0.008	Stratified
April	0.007	Stratified	October	0.005	Stratified
May	0.005	Stratified	November	0.031	Weakly stratified
June	0.004	Stratified	December	-	well-mixed

3.2 Conceptual Model for The Lake Water Balance

Lake water balance approach physically accounts for the components of outflow from the system, inflow to the system and the change in storage within the system over a period of time. The water balance method is normally used to estimate the net groundwater inflow (groundwater inflow - groundwater outflow) to the lake, provided the other water balance components are known.

The water balance equation for a lake can be written as

$$\Delta V = \text{inflow} - \text{outflow}$$

where, ΔV is the change in lake storage for a selected period of time. Incorporating different inflow and outflow components and rearranging, the equation becomes:

$$SS_I - SS_O = (E_O + S_O \pm \Delta V - (P_I + S_I + D_I) \qquad (3.7)$$

where, SS_I = sub-surface inflow to the lake [L^3/T]
SS_O = sub-surface outflow from the lake [L^3/T]
E_O = evaporation from the lake surface [L^3/T]
S_O = surface outflow from the lake [L^3/T]
ΔV = change in lake storage [L^3/T]
P_I = direct precipitation over the lake surface [L^3/T]
S_I = surface water inflow to the lake [L^3/T]
D_I = inflow to the lake through drains [L^3/T].

Water balance approach is used to determine the unmeasured components of a lake system for a particular condition. In the unsteady flow condition the changes in storage of the lake occur over a finite interval of time and therefore, the time interval used in the equation must be large. Further, all the known components should be estimated accurately for selected time interval. The magnitude of the computed component should be large relative to the sum of expected errors of in measurement of other components; otherwise, the error may overshadow the computed components.

Since the lake catchment is characterised by a multitude of faults and fractures coinciding with surface drainage system, the groundwater movement may preferentially be taking place along these zones. As the infiltration rates are higher in such zones, the surface runoff may be lower. The nature of the weathered zone indicates that the infiltrated water may move along preferential pathways towards the lake. The monthly spring discharge data and the chemistry of the springs indicate that the groundwater residence time may be short. Sukhatal sub-catchment of Nainital does not have any surface outflow and appears to be a closed type. The rainfall received in this sub-catchment is lost through infiltration and evaporation in a short time. Because of the proximity of Lake fault to Sukhatal lake, it is possible that most of the water is lost through underground seepage that subsequently recharges Nainital lake.

The change in lake storage responds quickly to direct precipitation and surface inflow as a consequence of steep hill slopes in the lake catchment. This is evident from daily lake level data. The other major component, that influences the lake storage, is the surface outflow. Evaporation from the lake surface during 24-hour time interval is negligible and is not readily discernible from daily lake level data. Apart from the above components that can be measured or estimated using standard methods, the lake level is also influenced by withdrawals from the lake.

Withdrawals from lake Nainital do not take place directly from the lake, but through the wells installed at the periphery of the lake. The entire quantum of water, that is pumped out of these wells may not completely be from lake seepage. However, as the wells are located in unconsolidated landslide debris and occur very close to the lake, it is possible that a major portion of the pumped water is replenished by the sub-surface outflow from lake Nainital as seepage.

The sub-surface outflow towards the downstream side of the lake may be through the fractures and faults. Seepage from lake may not be occurring through lakebeds as they are characterised by thick layer of fine sediments. Therefore, sub-surface outflow may be occurring mainly in the epilimnion zone. Outflow from the lake recharges the unconfined aquifer, which in turn, discharges through numerous springs located in the downstream side of the lake. Therefore, in the absence of groundwater level data in the unconfined aquifers at downstream side, the seepage from the lake could be quantified only from discharge measurements of springs that are hydraulically connected to the lake and located in the downstream side.

Keeping the above in view, the sub-surface outflow from the lake may be divided into two parts viz., withdrawal from the lake through the wells installed in the northern bank (Wo) and outflow to the springs (SPo). Therefore, Equation (3.7) may be modified by incorporating Wo and SPo in place of SSo.

The various flow components required for water balance study of the lake are shown in Figure 3.14.

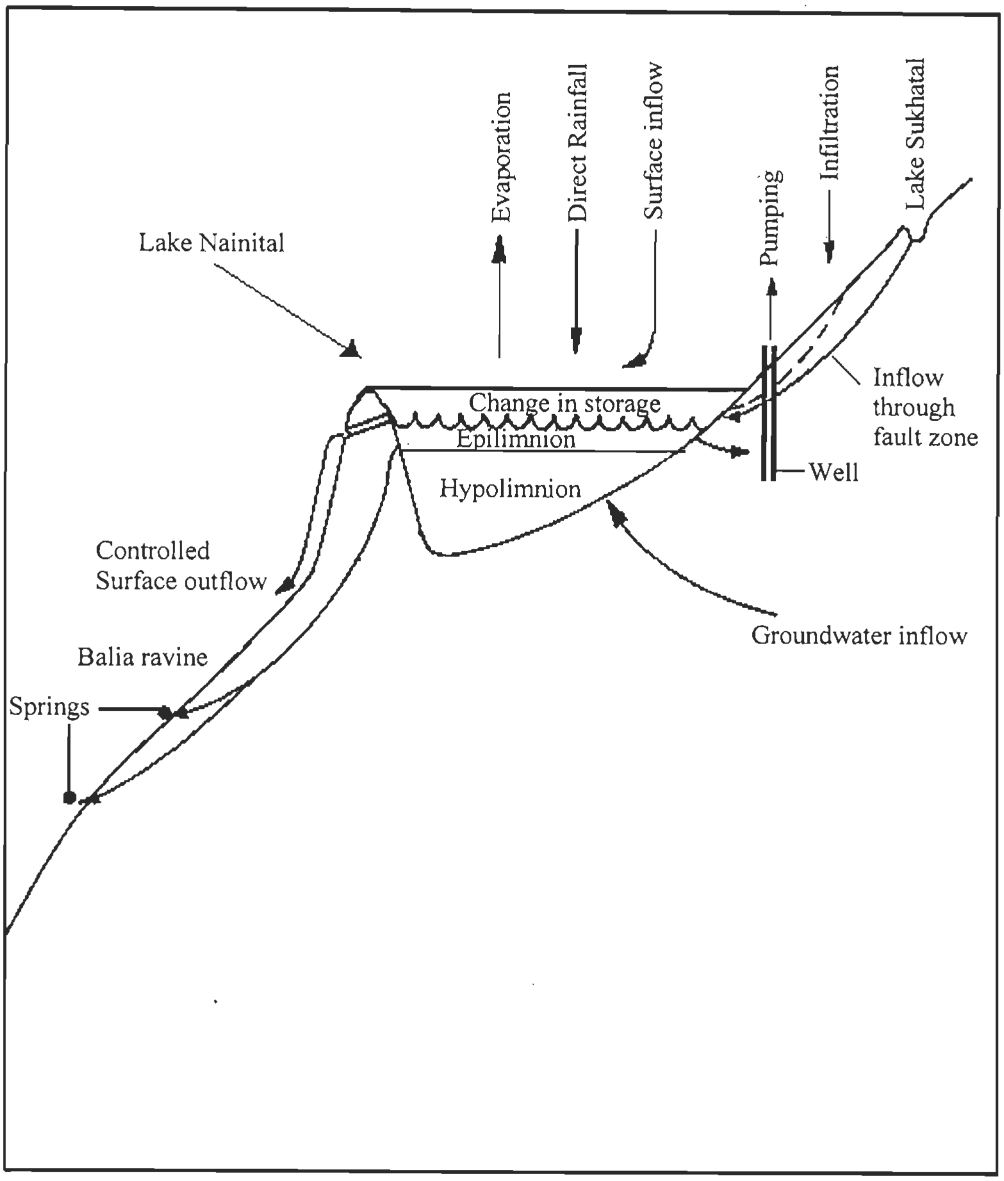


Figure 3.14 Conceptual model showing various water balance components of Nainital lake.

3.3 Water Balance of Lake Nainital

Water balance of lake Nainital has been computed for the period 1994 and 1995. The different methods adopted for estimating/computing those parameters that have been used in Equation (3.7), are presented and the results are discussed in the following sections.

3.3.1 Precipitation

The mean rainfall in the lake basin have been computed from rainfall data collected from four raingauge stations viz. Alma Cottage, ATI Campus, Snowview and Rattan Cottage using Theissen polygon method and their respective Theissen weights are 0.16, 0.45, 0.14 and 0.25. The computed mean monthly rainfall of Nainital basin is given in Table 3.14.

Table 3.14 Mean monthly rainfall (mm) of Nainital lake basin.

Months	1994	1995	Months	1994	1995
January	62.3	46.0	July	490.5	509.0
February	15.3	59.3	August	431.8	632.0
March	0.0	62.3	September	43.3	341.4
April	61.8	5.8	October	5.0	0.0
May	28.3	18.5	November	4.3	0.0
June	242.0	89.0	December	1.8	10.2

3.3.2 Surface Inflow

Surface inflow to the lake in response to rainfall has been estimated by two methods viz. Lake Level Trend Analysis (LLTA) and Soil Conservation Service - Curve Number (SCS-CN) methods.

3.3.2.1 Lake Level Trend Analysis method

In general, lake level follows a decreasing trend upto mid-June, and then follow an increasing trend upto mid-October. The lake level trend can be defined as the rate of change of lake level at any time. The drop in lake level indicates lesser inflow and higher outflow and conversely the rise in water level indicates higher inflow and lesser outflow from the lake. Figures 3.15a and 3.15b show the variation in daily lake level during the years 1994 and 1995 respectively in addition to daily rainfall during the corresponding periods. Lake level trend accounts for different losses from the lake including sub-surface outflow as well as groundwater

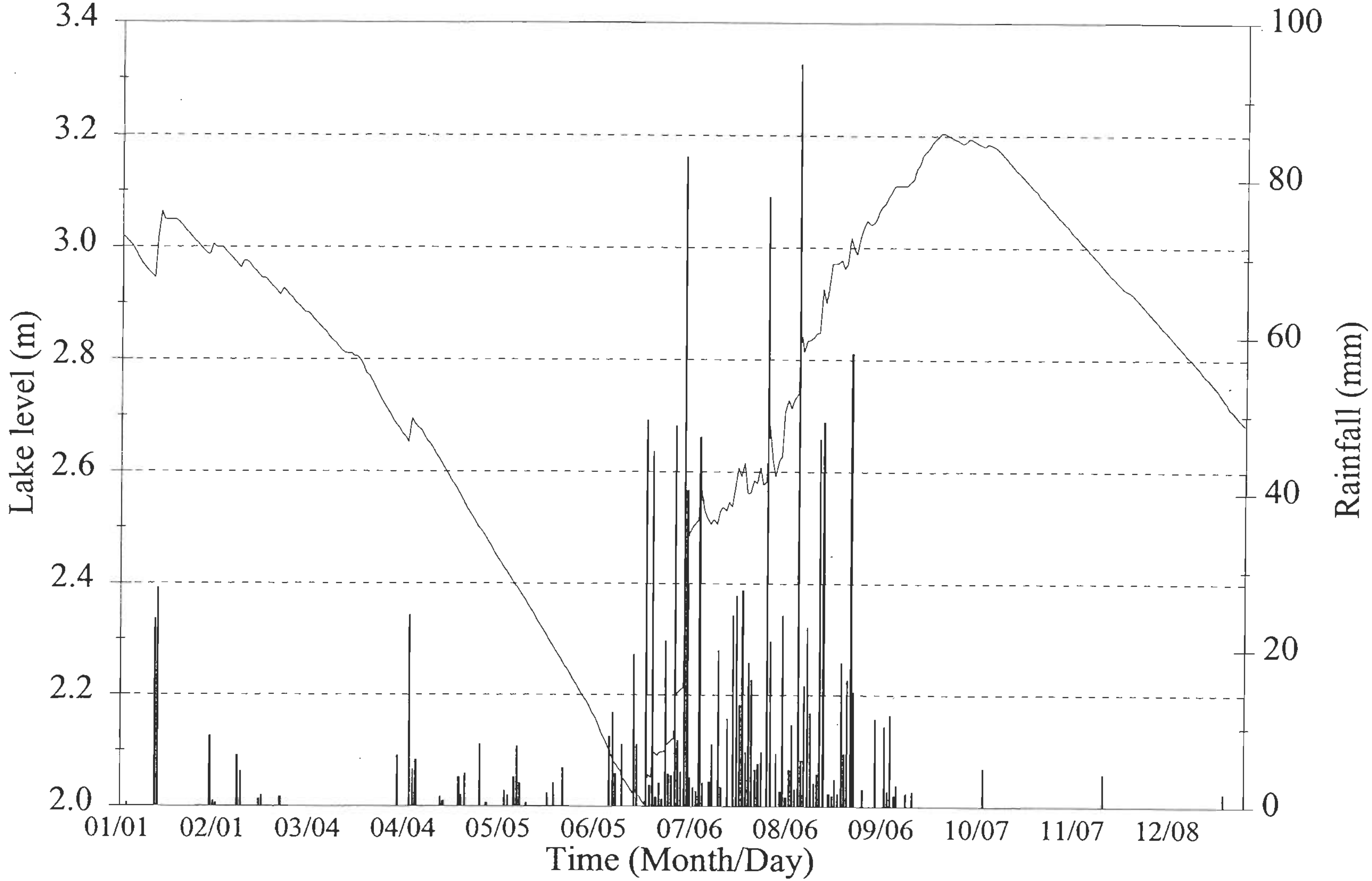


Figure 3.15a Variation of the lake level with daily rainfall during 1994.

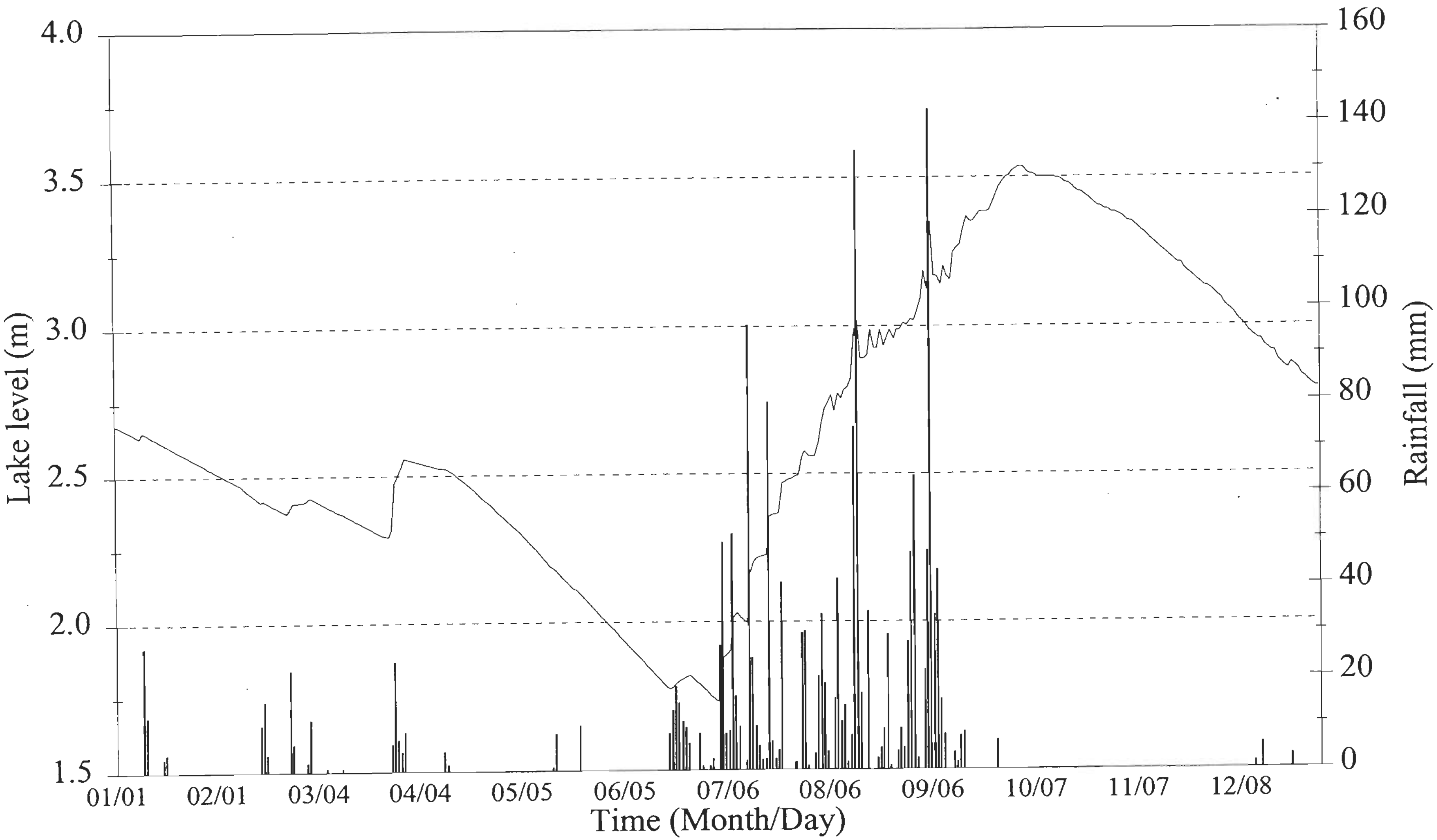


Figure 3.15b Variation of the lake level with daily rainfall during 1995.

inflow, as they do not vary drastically over shorter time intervals (24-hours). The LLTA method is similar to the well hydrograph analyses method used in the estimation of groundwater recharge. The equation used for estimation of the surface inflow by LLTA method is given below.

$$RO_i = \frac{\left[\frac{(OF_i - T_i - DRF_i)}{1000} + LA_i + O_i\right]}{CA} \tag{3.8}$$

Where,
RO_i = Estimated surface inflow (m)
OF_i = Observed fluctuation in the lake level (mm)
T_i = Lake level trend averaged over a seven day period (mm)
DRF_i = Direct rainfall falling on the surface area (mm)
LA_i = Surface area of the lake (m^2)
O_i = Surface outflow from the lake (m^3)
CA = Area of the lake catchment (m^2) = 3491707 m^2
Subscript i = time period (24 hours ending at 0830 hours)

3.3.2.2 Soil Conservation Service - Curve Number method

Surface inflow has also been estimated by Soil Conservation Service - Curve Number (SCS-CN) method (148). The widely accepted SCS-CN method is more useful for the estimation of 24 hour run-off (Q) from small catchments (135, 119):

$$Q = \frac{(P - 0.2S)^2}{P + 0.8S} \tag{3.9}$$

Parameter S depends upon the catchment characteristics, such as hydrologic soil type, land use and treatment, ground surface conditions and Antecedent Moisture Conditions (AMC). AMC refers to the soil water content at a given time. The SCS-CN method has three levels of AMC, depending on the total rainfall in the 5-day period preceding a storm, viz. AMC I, AMC II and AMC III. In AMC I, the soils are dry, but not to the wilting point. AMC II reflects average conditions and AMC III has highest run-off potential with the catchment practically saturated from antecedent rainfalls. S is expressed as a function of Curve Number (148):

$$S = \frac{1000}{CN} - 10 \tag{3.10}$$

CN is determined from the hydrologic soil type and antecedent moisture conditions. Appropriate CN for a variety of land uses, soil treatment or farming practices alongwith hydrological conditions (state of vegetation growth) have been presented for AMC II (148). The CN has been taken from the table of curve numbers for urban areas and agricultural lands by using land use information of the lake catchment (148, 119). Table 3.15 presents the type and percent area under different land use, appropriate CN and the composite CN for AMC II for Nainital lake catchment. The CN, thus determined for AMC II, has been converted to that of AMC I and AMC III using the following relations (67):

$$CN_I = \frac{CN_{II}}{2.3 - 0.013\ CN_{II}} \tag{3.11a}$$

$$CN_{III} = \frac{CN_{II}}{0.43 + 0.0057\ CN_{II}} \tag{3.11b}$$

Surface inflow to the lake has been estimated, with appropriate curve numbers for different AMC and daily rainfall data, by using Equations (3.9) and (3.10). Estimated runoff in the units of depth have been converted into units of volume using exclusive catchment area of the lake (3491707 m^2). Since Sukhatal sub-catchment does not contribute to surface inflow to the lake, it has been excluded in the estimations.

Table 3.15 Determination of composite Curve Number for AMC II based on land use data of Nainital lake catchment.

Land use	Ratio of area under given land use to the total catchment area	Hydrologic soil group	CN for AMC II
Reserved Forest	0.2352	A	30
Other Forest	0.2294	B	36
Built-up	0.4545	C	92
Barren	0.0448	C	86
Flats near lake	0.0125	A	76
Roads	0.0237	-	98
Composite Curve Number for AMC II			64

Daily surface inflow estimated by LLTA and SCS-CN methods for the year 1995, are shown in Figure 3.16. It is seen from the figure, that the values calculated by the two methods compare well. However, the minor differences are due to the influence of sub-surface inflow/ outflow on the lake level trend. The surface inflow estimated by semi-empirical SCS-CN method, is reliable, as it is comparable with the independent estimate made by another (LLTA) method. Since the SCS-CN method is a simpler and well established method for the estimation of surface inflow, the results of this method have been used in the lake water balance. Calculated surface inflow to the lake for the years 1994 and 1995 are presented in Tables 3.25a and 3.25b.

3.3.3 Inflow Through Drains

As discussed earlier, most of the drains are generally dry except during rainy season. However, Nainadevi and Rickshaw stand drains flow throughout the year. Measurements of discharge in these two drains were made at discrete time intervals by using a pygmy current meter. The flow rates have been estimated using the velocity-area method. Discharge of other drains, were also measured, during monsoon season. The total monthly discharges in the drains have been calculated from the measured flow rate and are presented in Tables 3.25a and 3.25b.

3.3.4 Change in Storage

Based on lake bathymetric data, four linear regression equations of the type $y = mx + c$, have been developed in order to calculate the lake surface area from lake level data observed in UPPWD lake level staff gauge installed near the sluice gates in Tallital end of lake Nainital. Depending on the lake level, appropriate equation may be used to calculate the lake surface area, which in turn may be used in the prismoidal formula for computing the change in lake storage. The values of regression coefficients m and c, and the validity range are presented in Table 3.16. The change in lake storage estimated from the lake level data for the years 1994 and 1995 are presented in Tables 3.25a and 3.25b.

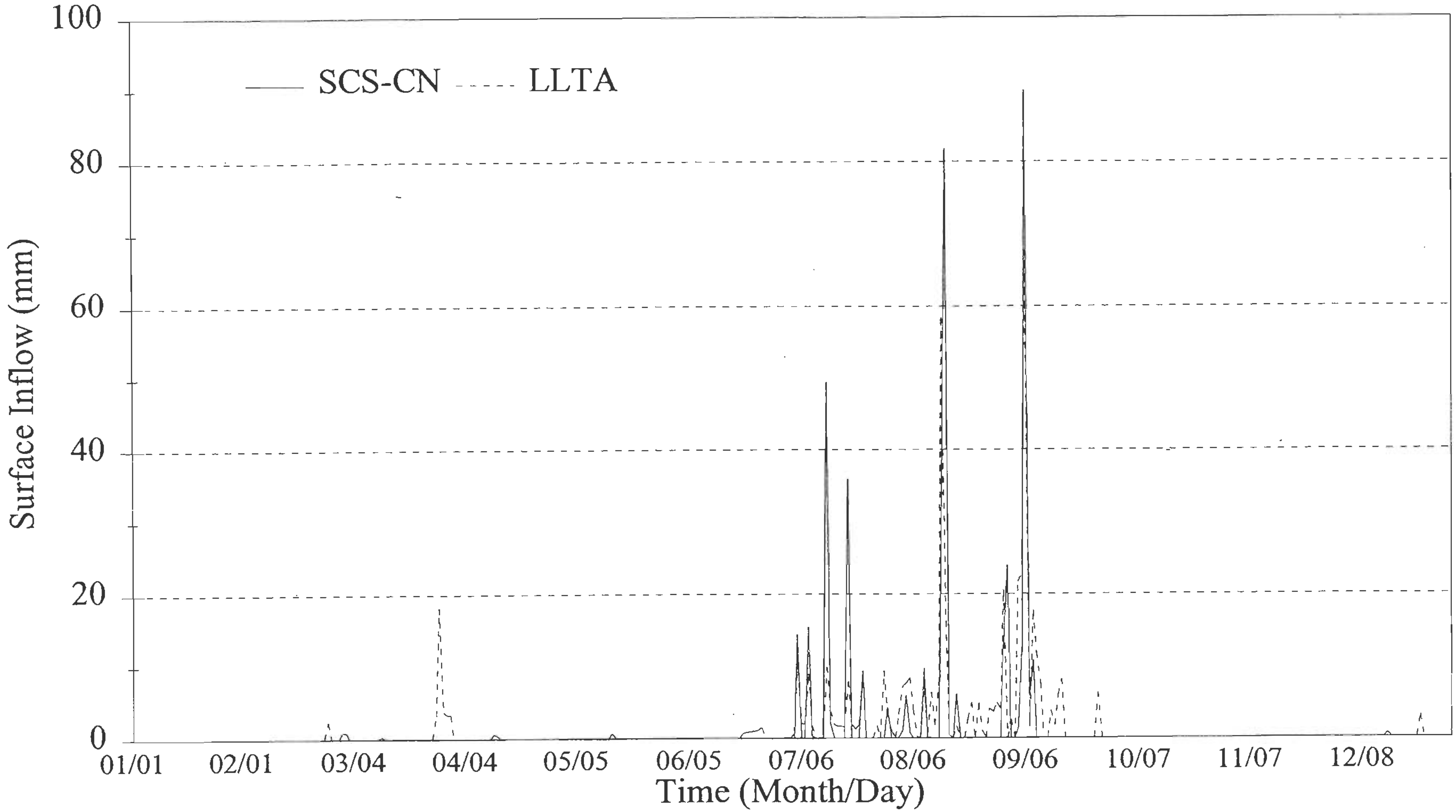


Figure 3.16 Comparison of daily variation in surface inflow to the lake in 1995 estimated by LLTA and SCS-CN methods

Table 3.16 Values of regression coefficients for the estimation of lake surface area (m^2) from lake level (mm) data.

For the Lake level in UPPWD gauge (validity range in mm)		slope (m)	intercept (c)
From	To		
3810	2441	6.86	437232
2440	1801	14.35	418999
1800	1491	6.12	433787
1490	0000	2.29	438698

3.3.5 Surface Outflow

Water level in the lake is maintained by local authorities, by regulating surface outflow through sluice openings at Tallital end. Discharge from the lake outlet has been calculated by using the following hydraulic equation (19):

$$Q = 252780 * (h_1 - h_2) * \frac{\sqrt{h_1 + h_2}}{2} \tag{3.12}$$

where, Q is the discharge in cubic feet per hour, h_1 and h_2 are respectively the heads to the bottom and top of the rectangular sluice opening in feet. h_1 and h_2 have been obtained from sluice gate operation records. In the calculation of surface outflow, leakage from sluice gates was also taken into consideration. Surface outflow estimated for the years 1994 and 1995 are presented in Tables 3.25a and 3.25b respectively.

To understand the overall change in the lake hydrological characteristics in the past 100 years, surface outflow has also been calculated from historical lake level and sluice operation records available with UPPWD for the period 1895-1995 . The results are shown in Figure 3.17 and the calculated surface outflow from the lake are presented in Table 3.17. The results indicate that surface outflow decreases for a given rainfall during the past three decades. The reduction, is mainly due to increased pumping from the northern banks to meet the domestic demands.

Table 3.17 Long-term annual rainfall and surface outflow of Nainital lake. Data from UPPWD, Nainital.

Year	Rain fall (cm)	Surface outflow (cm)	Year	Rainfall (cm)	Surface outflow (cm)	Year	Rainfall (cm)	Surface outflow (cm)
1895	247.9	1150	1931	174.2	123	1959	193.9	164
1896	176.3	504	1932	311.8	1440	1960	229.7	694
1898	258.7	1498	1933	246.9	236	1961	285.0	1358
1899	173.5	736	1934	268.0	1248	1962	267.2	518
1900	178.9	631	1935	171.1	98	1963	286.1	1377
1901	244.7	1335	1936	309.9	1527	1964	224.5	1205
1902	205.5	788	1937	312.4	1211	1965	164.3	46
1903	149.5	431	1938	255.1	1186	1966	197.0	519
1904	213.7	1583	1939	162.4	77	1967	268.8	1003
1905	184.2	1455	1941	213.0	123	1968	171.6	58
1906	251.8	2097	1942	287.9	1457	1969	262.5	811
1907	99.0	585	1943	231.4	786	1970	205.9	466
1908	170.4	749	1946	215.9	845	1971	254.9	1184
1909	305.8	2400	1947	246.8	760	1973	259.9	66
1910	368.9	3214	1948	291.5	846	1975	255.1	659
1911	199.6	1244	1949	311.8	1222	1977	230.9	459
1912	161.6	885	1950	279.4	1622	1978	220.3	763
1913	163.4	1139	1951	231.0	745	1989	194.3	569
1914	262.0	2188	1952	185.3	389	1991	142.6	146
1915	290.8	2025	1953	215.0	534	1992	160.5	310
1916	260.2	1980	1954	353.4	1491	1993	264.3	942
1922	289.4	1982	1955	250.4	614	1994	138.6	298
1923	203.3	1229	1957	252.5	1292	1995	172.7	403
1924	260.9	1366	1958	214.4	442			

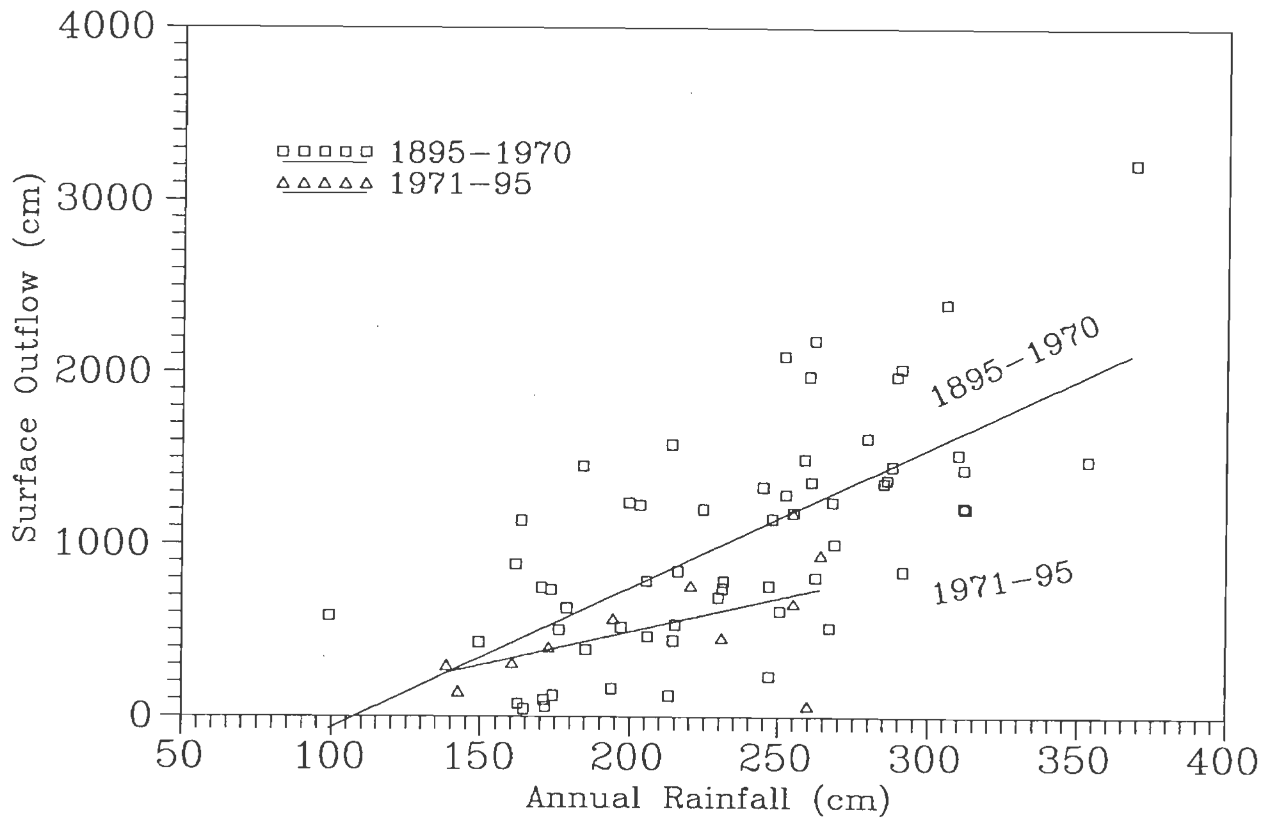


Figure 3.17 Variation of Surface outflow from the lake with the annual rainfall in the catchment.

3.3.6 Lake Evaporation

Estimation of evaporation is rather complicated. Of the three types of approaches available for the estimation of lake evaporation, viz. energy balance method, mass transfer method and Penman type combination method, the first one yields more reliable results, provided data on all energy terms are available (160). Alternatively, if sufficient data are available, the combination method yields better results. Therefore, for Nainital lake, the modified Penman method (75) has been chosen and the equation for the estimation of evaporation (E in mm) is:

$$E = \frac{\Delta}{\gamma + \Delta} \cdot (Q_N + G) + \frac{\gamma}{\gamma + \Delta} \cdot 15.36 \cdot (0.5 + 0.01 \cdot U_2) \cdot (e_0 + e_a) \qquad (3.13)$$

where, γ and Δ are the weighting factors. The following equation has been used for computation of Δ (mb/°C):

$$\Delta = \frac{25083}{(T + 237.3)^2} \cdot \exp\left(\frac{17.3 \cdot T}{T + 237.3}\right) \qquad (3.14)$$

where, T is the mean air temperature in °C. The psychrometric constant (γ) has been computed using the following equation:

$$\gamma = \frac{0.61 \cdot P}{1000} \qquad (3.15)$$

where, P is the mean atmospheric pressure in mb. Atmospheric pressure data collected from UPSO have been used in the present study.

Q_N is the net solar radiation. Since the solar radiation data was not available, the net radiation has been computed using the following relationship:

$$Q_N = Q_S \cdot (1 - \alpha) - Q_{LN} \qquad (3.16)$$

where, α is the albedo (reflection coefficient), Q_S is the global solar radiation and Q_{LN} is the net long wave radiation. α for the present study has been computed by using values of different amounts of cloud cover (143). Sunshine hours data collected by UPSO have been used to generate the cloud cover data. The following equation based on IMD data for whole of India, has been used for computing Q_S (121):

$$Q_S = Q_A * [0.325 * COS\Phi + 0.385 * (\frac{n}{N})] \tag{3.17}$$

where Φ is the latitude and n and N are actual and maximum possible sunshine hours respectively. The extra-terrestrial radiation (Q_A) and N for the site have been computed using the Duffie-Beckman equations (2). Q_{LN} has been computed by the following equation (132):

$$Q_{LN} = - f \epsilon' \sigma T^4 \tag{3.18}$$

where, σ is Stefan-Boltzmann constant, T is absolute temperature in degree Kelvin, f is adjustment factor for cloud cover, which is roughly equal to (0.1+0.9*n/N) (135) and ϵ' is emissivity factor, which has been computed using the Idso-Jackson equation (106). U_2 is wind velocity in m/s at 2 m above surface. e_0 and e_a are saturated vapour pressure (mb) at the water surface temperature and actual vapour pressure (mb) at air temperature respectively. e_0 can be calculated using the following equation (94):

$$e_0 = 33.8639 * [(0.00738\ T_0 + 0.8072) - 0.000019 * |1.8\ T_0 + 48| + 0.001316] \tag{3.19}$$

Actual vapour pressure, e_a, is computed by multiplying relative humidity with the saturated vapour pressure at air temperature. For Penman equation, the air and water temperatures and relative humidity were measured at the lake site. The estimated and measured values of the parameters used in the computation of lake evaporation are presented in Table 3.18. Monthly evaporation loss from the lake for the years 1994 and 1995 are presented in Tables 3.25a and 3.25b respectively.

Table 3.18 Values of measured and estimated parameters used in the estimation of evaporation from Nainital lake by modified Penman method for the year 1995.

Month	Measured						Estimated	
	Atm. Press. (mb)	Sun Shine (Hrs)	Air Temp. (°C)	Lake Temp. (°C)	Wind Velo. (m/s)	Rel. Humi. (%)	Alpha Coeff. -	Net Radn. (ly/d)
Jan	812.78	6.76	4.43	8.82	3.19	58.42	0.087	145.19
Feb	812.37	6.22	6.41	8.78	3.23	53.42	0.074	191.07
Mar	813.48	8.61	10.05	12.08	4.89	47.96	0.068	278.47
Apr	812.82	6.47	15.31	15.96	4.15	47.30	0.058	310.60
May	812.26	8.25	20.22	19.90	4.73	61.95	0.061	368.75
Jun	809.57	2.41	21.78	22.47	3.60	68.88	0.056	288.31
Jul	809.27	3.10	18.77	21.51	2.75	85.74	0.057	295.06
Aug	810.02	2.91	18.41	20.93	2.72	87.43	0.056	272.73
Sep	811.97	4.53	17.18	20.13	2.25	81.10	0.064	256.56
Oct	814.50	8.28	15.32	18.42	2.12	72.64	0.062	239.31
Nov	813.86	8.94	12.22	14.64	1.75	51.17	0.076	169.25
Dec	814.15	7.76	8.67	11.17	1.82	53.81	0.087	129.19

3.3.7 Estimation of Proportion of Lake Water during Pumping

The proportion of the lake water being pumped from the wells located near the lake was estimated by isotopic tracer technique. A two-component mixing model, as discussed earlier in Chapter 2.0, has been applied. Groundwater isotope index has been calculated from the data of upstream springs given in Table 3.40, and the lake isotopic index has been calculated from volume-weighted averages using data given in Table 3.38. The $\delta^{18}O$ data of admixture i.e., the well water, end-member indices alongwith the proportion of lake water being pumped are presented in Table 3.19. The results show that proportion of lake water component in the water pumped from the wells is lower in non-monsoon seasons, as compared with monsoon season.

Table 3.19 Proportion of lake water in the well water being pumped (Wo) along with $\delta^{18}O$ of end-members and admixture.

Month	$\delta^{18}O$ (‰)			Proportion of lake water (%)
	Lake	Groundwater	Well	
February, 1995	-7.3	-8.2	-8.0	25
March, 1995	-7.1	-7.5	-7.4	25
May, 1995	-7.1	-7.5	-7.4	30
August, 1995	-6.3	-8.9	-6.8	80
November, 1995	-8.2	-7.9	-8.0	40

3.3.8 Interconnection between Nainital Lake and Downstream Springs

As discussed earlier, it is likely that some of the downstream springs may be related to the lake. In order to understand the inter-relationship between the lake and the springs, hydrochemistry and isotopic characteristics have been studied. The hydrochemical and isotopic data are presented in Tables 3.20a, 3.20b, 3.21 and 3.22 and plotted in Figures 3.18, 3.19 and 3.20.

Table 3.20a SO_4^{2+} and Cl^- data of the Nainital lake and surrounding springs during winter 1994-95.

Location	Month	SO_4^{2+} (mg/L)	Cl^- (mg/L)	Location	Month	SO_4^{2+} (mg/L)	Cl^- (mg/L)
Lake - Tallital top	Dec., 94	140.0	6	S-12 spring.	Dec., 94	408	2
Lake - Tallital bottom	Dec., 94	107.0	8	Sipahidhara spring	Dec., 94	116	6
Lake - Mallital top	Dec., 94	139.0	4	Gupha M.T. spring	Dec., 94	119	6
Lake - Mallital bottom	Dec., 94	124.0	8	Rais Hotel spring	Mar., 95	117	8
S-3 spring	Dec., 94	197.0	8	Sariyatal spring.	Mar., 95	149	16
S-9 spring	Dec., 94	132.0	2				

Table 3.20b Total cation (TZ^+) data of samples collected from Nainital lake and surrounding springs during 1995-96. All values are in equivalents per million.

Month	Lake (mean)	Lakeview Sp.	Open well	Sipahidhara	S-3 Sp.
March, 95	7.37	11.43	-	7.57	7.99
May, 95	6.94	7.54	6.49	6.74	8.44
June, 95	6.95	-	-	6.96	-
August, 95	6.76	9.03	11.74	7.03	3.92
September, 95	7.01	11.11	10.13	6.93	4.58
November, 95	8.50	11.76	8.33	7.89	6.48
December, 95	7.64	9.98	8.97	6.76	-
January, 96	8.43	10.81	9.91	7.12	-

Table 3.21 $\delta^{18}O$ values of downstream springs. Samples collected during December, 1994.

Spring ID	Altitude (m above m.s.l.)	$\delta^{18}O$ (‰)	Spring ID	Altitude (m above m.s.l.)	$\delta^{18}O$ (‰)
S1	1850	-10.6	S7	1640	-11.8
S2	1790	-7.5	S9	1760	-10.7
S3	1730	-7.4	S10	1760	-10.2
S4	1720	-7.0	S11	1650	-10.9
S5	1730	-7.7	S12	1700	-10.7
S6	1750	-11.0	Gupha M.T	1785	-9.5

Sulphate versus chloride plots of Nainital lake and downstream Sariyatal and Balia ravine springs show clustering (Figure 3.18). Other downstream springs, such as those in Kailakhan area and Durgapur, show enrichment in sulphate content as compared to the lake. When total cations (TZ^+) of different springs are normalised to those of the lake, Sipahidhara spring shows little variation with time as compared to upstream springs and wells (Figure 3.19). This indicates that the lake is the main source for Sipahidhara spring. Likewise, the isotopic values of some of the springs are very much similar to the lake. $\delta^{18}O$ of Gupha Mahadev spring and the lake are -9.5‰

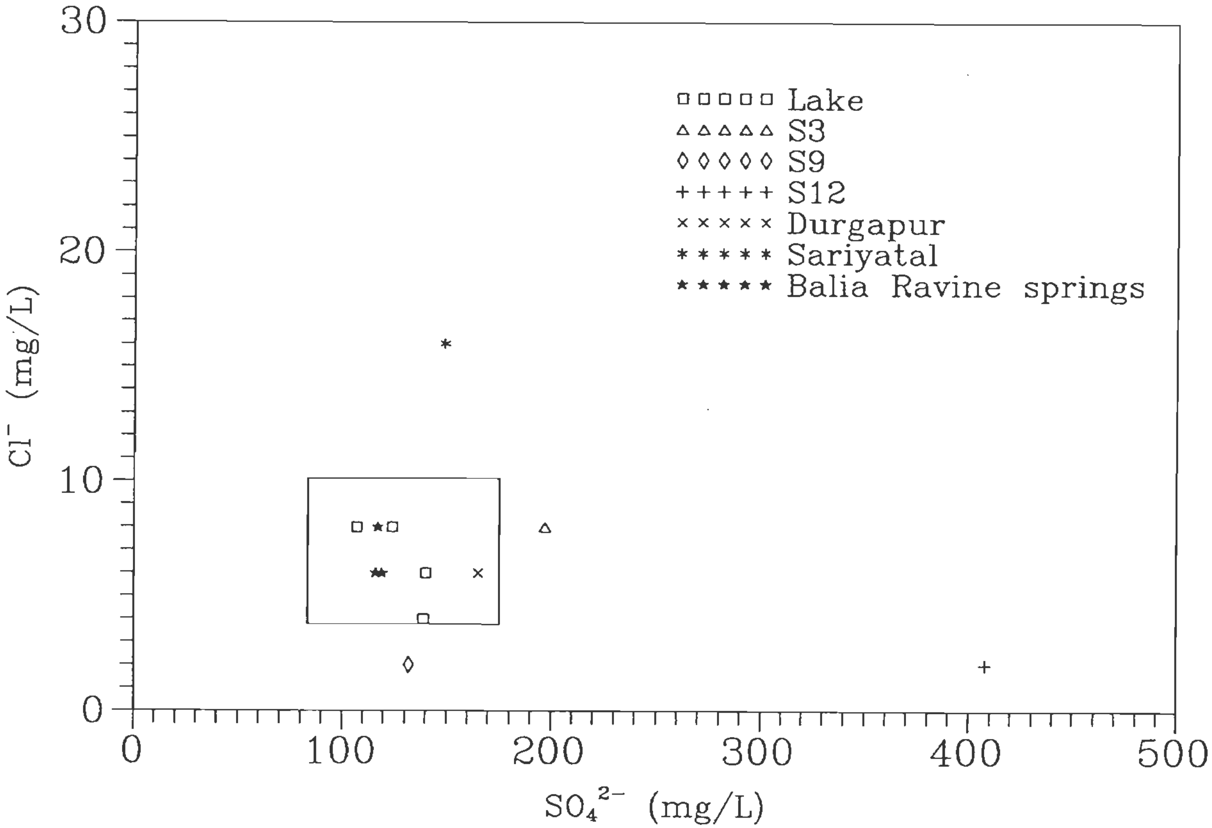


Figure 3.18 Sulphate – Chloride cross plot for different water sources in Nainital area.

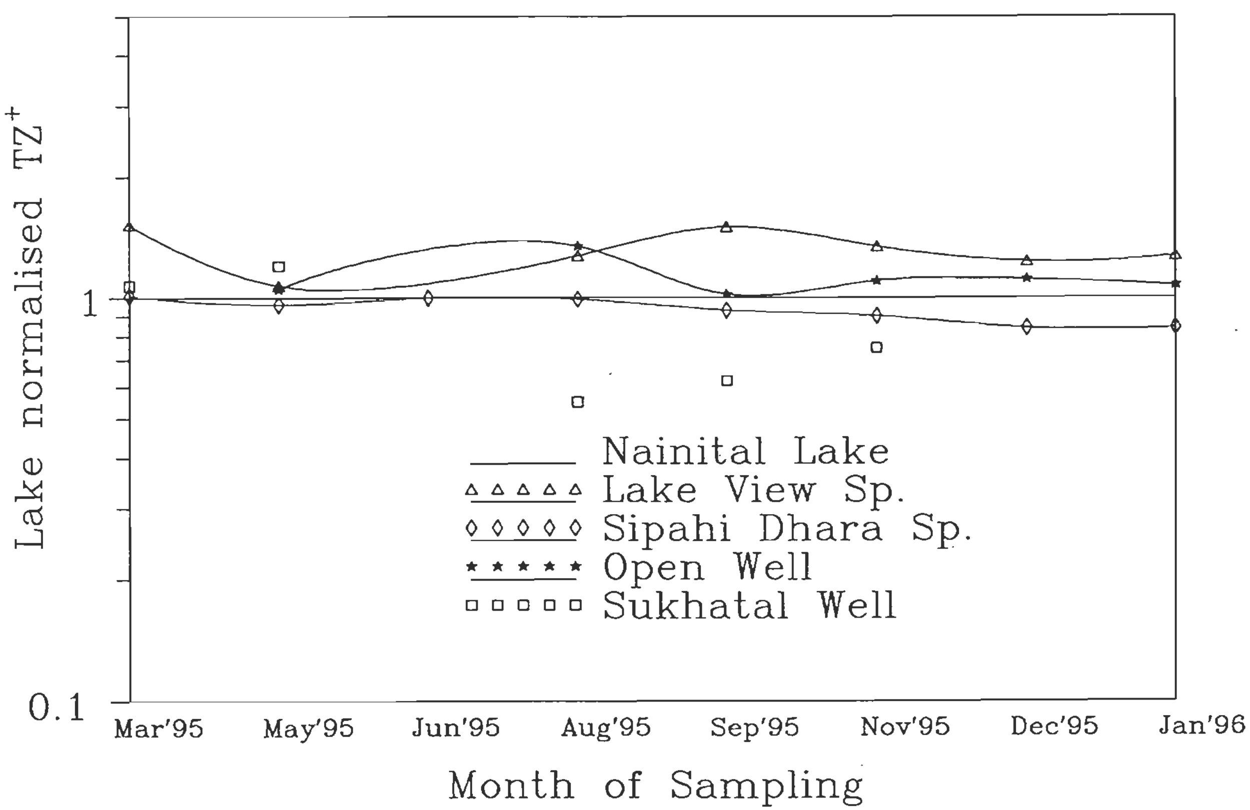


Figure 3.19 Variation of total cations (TZ^+) of different water sources with time.

and -9.6‰ respectively. Springs located in Kailakhan area show relatively heavier $\delta^{18}O$ (-7.0 to -7.5 ‰) during winter, which is comparable to $\delta^{18}O$ of the lake in summer. However, some of the springs such as S1, S6, S7, S9, S10, S11 and S12 show depleted $\delta^{18}O$ values of -10.2 ‰ to –11.8 ‰ as compared to those of the lake (-5.0‰ to –9.6‰).

Further, samples were collected from Gupha Mahadev Temple Spring during different seasons to establish the nature of variation in $\delta^{18}O$ in relation to that of Nainital lake. The results of this spring are given in Table 3.22.

Table 3.22 $\delta^{18}O$ and δD of Gupha Mahadev Temple spring during different seasons.

Month	$\delta^{18}O$ (‰)	δD(‰)	Month	$\delta^{18}O$ (‰)	δD(‰)
December,94	-9.5	-64	August, 95	-7.1	-51
February, 95	-7.9	-55	September, 95	-8.8	-57
March, 95	-6.8	-44	November, 95	-8.7	-48
May, 95	-5.6	-51	September, 96	-8.5	-56
June, 95	-	-50			

Large variation in $\delta^{18}O$ values of Gupha Mahadev Temple Spring during different seasons indicate that the spring is replenished from the epilimnion zone of the lake (Figure 3.20). The stable isotope data confirms the inference (drawn from the comparison of ion concentrations of the Sipahidhara and Gupha Mahadev springs to the vertical concentration profile of the lake) that these springs receive water from the epilimnion zone of the lake.

3.3.9 Estimation of Sub-surface Outflow Through Springs

Average monthly discharge of nine downstream springs located in Balia ravine were monitored by UPPWD for the period of 1948-52. The data suggest that out of the nine springs, only Rais Hotel and Sipahidhara springs account for about 92% of the aggregate discharge. Presently, many of these springs are dry, and the discharge of Sipahidhara is considerably reduced. The total monthly discharge of all the springs and that of Sipahidhara spring measured during the period 1948- 1952 alongwith discharge of Sipahidhara spring monitored during 1995 are presented in Table 3.23.

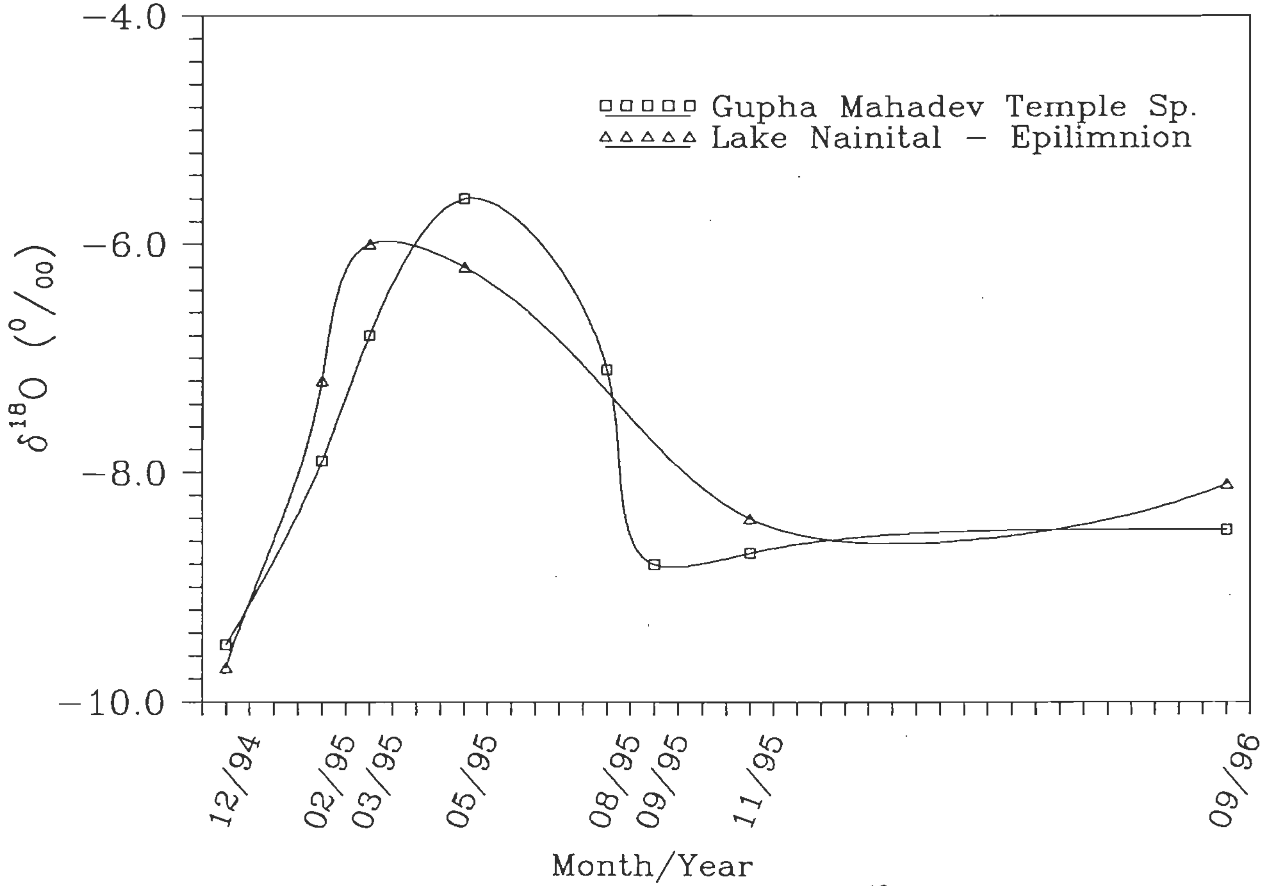


Figure 3.20 Temporal variation in $\delta^{18}O$ of Gupha Mahadev Spring and Lake epilimnion zone

Table 3.23 Comparison of discharge (x10^3 m^3) of Sipahidhara spring (SD) with that of all the springs.

Year	1948		1949		1950		1951		1952		1995
	SD	Total	SD	Total	SD	Total	SD	Total	SD	Total	SD
Jan	88	165	135	260	88	171	87	178	87	178	14
Feb.	79	148	109	199	80	156	79	156	79	156	11
Mar .	89	169	108	195	90	171	89	172	89	172	1
Apr.	84	159	83	160	80	157	81	152	81	152	0.6
May	75	150	75	138	75	146	33	89	33	87	-
Jun.	63	125	73	142	71	140	31	82	30	82	-
Jul.	70	154	300	539	-	36	302	424	109	223	-
Aug	301	554	301	549	401	709	401	583	201	361	40
Sep.	315	588	389	671	291	535	290	530	146	316	50
Oct.	302	570	241	446	235	443	235	461	139	293	28
Nov	246	425	166	327	164	328	164	324	124	253	19
Dec.	172	309	152	256	156	263	155	260	116	233	18

From the above data, it is seen that there is a reduction of about 85% in the discharge of Sipahidhara spring in the past 50 years. The reduction is probably due to clogging of subterranean pathway as a consequence of lake sedimentation (89). Mean monthly ratio of discharge of Sipahidhara spring to the total discharge of all the springs have been calculated from the data for the period 1948-52. Total discharge of all the springs located in the Balia ravine, has been computed for the years 1994 and 95 by using the calculated ratio and the discharge of Sipahidhara spring observed in 1995, The results are presented in Tables 3.25a and 3.25b.

3.3.10 Uncertainties in the Estimation of Water Balance Components

Overall accuracy of the water balance method depends on the accuracy of each flow component. Lake water balances, which are determined without error estimations could be misleading (158). In case of certain components of water balance, the amount of error could not be correctly evaluated due to the nature of the estimation methods. In the present study, errors in the estimation of different components of water balance are assumed to be independent and normally distributed. Estimated standard errors associated with different water balance components are presented in Table 3.24.

Table 3.24 Estimated standard errors for different components of the lake water balance

Component	Symbol	Method of determination	Estimated standard error	Remarks
Change in Storage	ΔV	Prismoidal formula using lake level data	10%	
Direct Rainfall	RF_I	Theissen polygon method using data from four rain gauge stations	10%	(158)
Surface Inflow	S_I	SCS-CN and LLTA methods	20%	
Inflow through drains	D_I	Discharge measurements at discrete time intervals	15%	(158)
Evaporation loss	E_o	Modified Penman method	15%	(158)
Surface outflow	S_O	Empirical formula	upto 5%	Varies
Outflow through springs	Sp_O	Discharge measurements at discrete time intervals	15%	
Outflow through wells	W_O	Pumping records and results of tracer tests	10%	
Subsurface Inflow	SS_I	Water balance equation	upto 10%	Estimated using Equation (3.21)

3.3.11 Estimation of Sub-surface Inflow

Calculated values alongwith standard error for each component of monthly water balance for the years 1994 and 1995 are presented in Tables 3.25a and 3.25b. Groundwater inflow to the lake (SS_I) and the standard error in the estimation of groundwater inflow (σ_{SSI}) have been calculated by the following equations:

$$SS_I = (E_O + S_O + W_O + SP_O + \Delta V) - (P_I + S_I + D_I) \qquad (3.20)$$

$$\sigma_{SSI} = [\sigma_{EO}^2 + \sigma_{SO}^2 + \sigma_{WO}^2 + \sigma_{SPO}^2 + \sigma_{\Delta V}^2 + \sigma_{PI}^2 + \sigma_{SI}^2 + \sigma_{DI}^2]^{1/2} \qquad (3.21)$$

It is seen from the results that the absolute error in groundwater inflow varies for monthly estimates, but for the annual estimates of groundwater inflow to the lake it is around 10%. The errors do not mask the results and therefore, the results may be considered as reliable.

Table 3.25a Estimates of different water balance components ($x10^3$ m^3) of Nainital lake for the year 1994 alongwith standard error in the estimates.

Months	ΔV	P_I	D_I	S_I	S_o	E_o	W_o	SP_o	SS_I
January	-8 ± 0.8	29 ± 2.9	54 ± 8.1			30 ± 4.5	61 ± 6.1	34 ± 5.1	34 ± 12.6
February	-53 ± 5.3	7 ± 0.7	53 ± 8.0			28 ± 4.2	61 ± 6.1	43 ± 6.5	19 ± 13.7
March	-95 ± 9.5		48 ± 7.2			50 ± 7.5	74 ± 7.4	46 ± 6.9	27 ± 17.3
April	-96 ± 9.6	28 ± 2.8	41 ± 6.2			64 ± 9.6	84 ± 8.4	47 ± 7.1	30 ± 18.7
May	-126 ± 12.6	13 ± 1.3	40 ± 6.0			73 ± 11	138 ± 13.8	52 ± 7.8	84 ± 23.8
June	9 ± 0.9	109 ± 10.9	40 ± 6.0	56 ± 11.2		62 ± 9.3	141 ± 14.1	48 ± 7.2	55 ± 24.9
July	174 ± 17.4	223 ± 22.3	75 ± 11.3	417 ± 83.4	427 ± 14.2	54 ± 8.1	322 ± 32.2	64 ± 9.6	326 ± 96.3
August	206 ± 20.6	197 ± 19.7	121 ± 18.2	354 ± 70.8	569 ± 27.7	47 ± 7.1	214 ± 21.4	117 ± 17.6	481 ± 87.9
September	67 ± 6.7	20 ± 2	104 ± 15.6		389 ± 42.8	51 ± 7.7	89 ± 8.9	127 ± 19.1	599 ± 51.3
October	64 ± 6.4	2 ± 0.2	75 ± 11.3		185 ± 20.9	47 ± 7.1	128 ± 12.8	65 ± 9.8	412 ± 30.2
November	-80 ± 8.0	2 ± 0.2	62 ± 9.3			29 ± 4.4	134 ± 13.4	74 ± 11.1	93 ± 21.7
December	-91 ± 9.1	1 ± 0.1	59 ± 8.9			29 ± 4.4	136 ± 13.6	66 ± 9.9	80 ± 21.5

Table 3.25b Estimates of different water balance components (x10^3 m^3) of Nainital lake for the year 1995 alongwith standard error in the estimates.

Months	ΔV	P_I	D_I	S_I	S_o	E_o	W_o	SP_o	SS_I
January	-75 ± 7.5	21 ± 2.1	54 ± 8.1			32 ± 4.8	112 ± 11.2	34 ± 5.1	28 ± 17.3
February	-38 ± 3.8	27 ± 2.7	53 ± 8.0			37 ± 5.6	89 ± 8.9	43 ± 6.5	51 ± 15.4
March	56 ± 5.6	28 ± 2.8	48 ± 7.2			56 ± 8.4	102 ± 10.2	46 ± 6.9	184 ± 17.7
April	-99 ± 9.9	3 ± 0.3	41 ± 6.2			62 ± 9.3	120 ± 12.0	47 ± 7.1	86 ± 20.4
May	-162 ± 16.2	8 ± 0.8	40 ± 6.0			72 ± 10.8	150 ± 15.0	52 ± 7.8	64 ± 26.5
June	-104 ± 10.4	39 ± 3.9	40 ± 6.0			57 ± 8.6	140 ± 14.0	48 ± 7.2	62 ± 21.9
July	367 ± 36.7	228 ± 22.8	75 ± 11.3	458 ± 91.6	40 ± 1.7	55 ± 8.3	353 ± 35.3	64 ± 9.6	118 ± 108.6
August	223 ± 22.3	289 ± 28.9	121 ± 18.2	610 ± 122.0	775 ± 26.0	51 ± 7.7	257 ± 25.7	117 ± 17.6	403 ± 135.1
September	221 ± 22.1	157 ± 15.7	104 ± 15.6	423 ± 84.6	998 ± 49.1	47 ± 7.1	97 ± 9.7	127 ± 19.1	806 ± 105.1
October	-71 ± 7.1		75 ± 11.3		212 ± 29.2	45 ± 6.8	94 ± 9.4	65 ± 9.8	270 ± 35.5
November	-127 ± 12.7		62 ± 9.3			35 ± 5.3	118 ± 11.8	74 ± 11.1	38 ± 23.2
December	-142 ± 14.2	5 ± 0.5	59 ± 8.9			26 ± 3.9	166 ± 16.6	66 ± 9.9	52 ± 25.9

3.4 Chemical Mass Balance

3.4.1 Hydrochemistry of Rainfall

Information pertaining to the rainfall chemistry is useful to derive Evapotranspiration Index (ET_{index}), i.e. the water loss by evapotranspiration during infiltration and percolation processes. Since no data regarding the chemistry of rainfall for Nainital lake catchment are available, the data pertaining to Lucknow (65) and Srinagar (96) stations have been used for the purpose. The calculated average concentrations of the major elements in the monsoon rainfall of Nainital region are (in mg/L): Ca^{2+} 4.44, Na^{+} 1.02, K^{+} 0.51, and Cl^{-} 0.99, which are within the range of values in continental rains (14).

3.4.2 Hydrochemistry of Surface Inflow and Inflow Through Drains

In order to understand the water chemistry of drains, samples were collected from different drains during different periods (Table 3.26) and analysed for major ions.

Table 3.26 Details of samples collected from drains

Month	No. of Drains	Month	No. of Drains
Feb.,94	1	Aug.,95	7
May, 94	1	Sep.,95	6
Oct.,94	11	Nov.,95	2
Dec.,94	1	Jan.,96	2
Feb.,95	1	Apr.,96	2
Mar.,95	2	Jul.,96	1
May, 95	2		

Water samples were collected during most of the sampling periods from the perennial Nainadevi and/or Rickshaw stand drains, but only during monsoon season from the seasonal drains located in Sher-ka-Danda hill. Chemical data of drains, sampled during monsoon and non-monsoon seasons are presented in Tables 3.27 and 3.28 respectively. The mean values are plotted in Figures 3.21a and 3.21b.

Table 3.27 Hydrochemistry of drain water samples collected during monsoon season from Nainital lake catchment. All concentrations in mg/L, E.C. in μS/cm.

Month	Location	pH	E.C.	HCO_3^-	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+
Sep./ Oct-94	RSD	-	-	350	45	26	86.6	44.7	-	-
	D - # 4	-	-	296	139	11	52.1	59.8	-	-
	D - # 11	-	-	242	152	14	52.9	70.5	-	-
	D - # 13	-	-	280	158	15	43.3	43.2	-	-
	D - # 16	-	-	290	85	13	43.3	72.9	-	-
	ND	-	-	288	158	-	33.7	81.6	-	-
Aug.- 95	RSD	7.4	909	278	144	42	36.9	71.9	27.0	11.9
	NDD	8.2	916	292	147	10	34.5	75.3	12.0	3.7
	D - # 4	8.1	775	346	125	18	15.2	110.3	20.0	5.9
	D - # 6	8.1	663	326	199	18	36.9	87.0	26.0	12.0
	D - # 10	7.7	807	288	158	46	40.1	77.3	31.0	11.1
	D - # 20	7.8	802	306	134	32	33.3	81.4	36.6	12.6
Sep.- 95	RSD	8.2	478	306	125	24	31.3	93.8	34.5	10.1
	D - # 4	-	-	290	136	30	35.3	76.3	17.1	4.6
	D - # 10	-	-	354	163	24	48.9	84.1	23.3	7.9
	D - # 13	-	-	306	137	26	36.1	74.8	18.2	5.0
	D - # 20	-	-	362	171	30	36.9	94.3	27.4	8.7
	NDD	-	-	310	117	6	80.2	50.5	-	-
Number of samples				18	18	17	18	18	11	11
Minimum value				242	45	6	15.2	43.2	12.0	3.7
Maximum value				362	199	46	86.6	110.3	36.6	12.6
Mean concentration				306	139	23	43.2	75.0	24.8	8.5

* NDD - Nainadevi drain; RSD Rickshaw Stand Drain.

Table 3.28 Hydrochemistry of drain water samples collected during non-monsoon seasons from Nainital catchment. All concentations in mg/L, E.C. in µS/cm.

Month	Location	pH	E.C.	HCO_3^-	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+
Feb.- 94	NDD	8.1	455	290	-	3	69.0	60.0	-	-
May- 94	NDD	7.5	1040	202	173	9	86.2	79.5	10.4	9.3
Dec.- 94	NDD	7.6	822	328	-	4	71.0	62.0	12.2	7.4
Feb.- 95	NDD	7.0	490	-	125	22	20.1	72.9	15.2	6.4
Mar.- 95	RSD	-	1147	350	189	44	71.9	74.4	27.6	16.0
	NDD	-	999	366	194	27	57.5	77.9	19.0	8.7
May- 95	RSD	7.1	1549	360	126	82	38.5	99.1	72.0	35.0
	NDD	7.1	1054	234	150	45	49.7	84.6	32.0	16.0
Nov.- 95	RSD	7.2	821	297	127	35	83.0	74.6	41.8	8.7
	NDD	7.0	796	288	133	16	88.6	66.6	23.1	5.8
Jan.- 96	RSD	-	-	298	94.5	24	47.3	93.3	9.6	6.6
	NDD	-	-	324	143	18	77.8	66.1	7.6	7.6
Apr.- 96	RSD	-	-	362	139	46	92.2	60.8	-	-
	NDD	-	-	392	108	56	38.5	83.4	-	-
Number of samples				13	12	14	14	14	11	11
Minimum value				202	95	3	20.1	60.0	7.6	5.8
Maximum value				392	194	82	92.2	99.1	72.0	35.0
Mean concentration				313	141	31	63.7	75.4	24.6	11.6

* NDD - Nainadevi drain; RSD Rickshaw Stand Drain.

It is seen that in general, the drain water show cation concentrations in the following order: $Mg^{2+} > Ca^{2+} > Na^+ > K^+$ and anion concentrations: $HCO_3^- > SO_4^{2-} > Cl^-$. Based on Stiff pattern diagrams, the drain water may be classified as Magnesium Bicarbonate type. Further, the data indicate that there is no significant seasonal variation in the mean concentration of different ions, except for K^+ and Cl^-. However, maximum concentration of Na^+, K^+ and Cl^- are higher during the non-monsoon season (72.0, 35.0 and 82 mg/L respectively) than in the monsoon season (36.6, 12.6 and 46 mg/L respectively). The variation may be due to disposal of sewerage

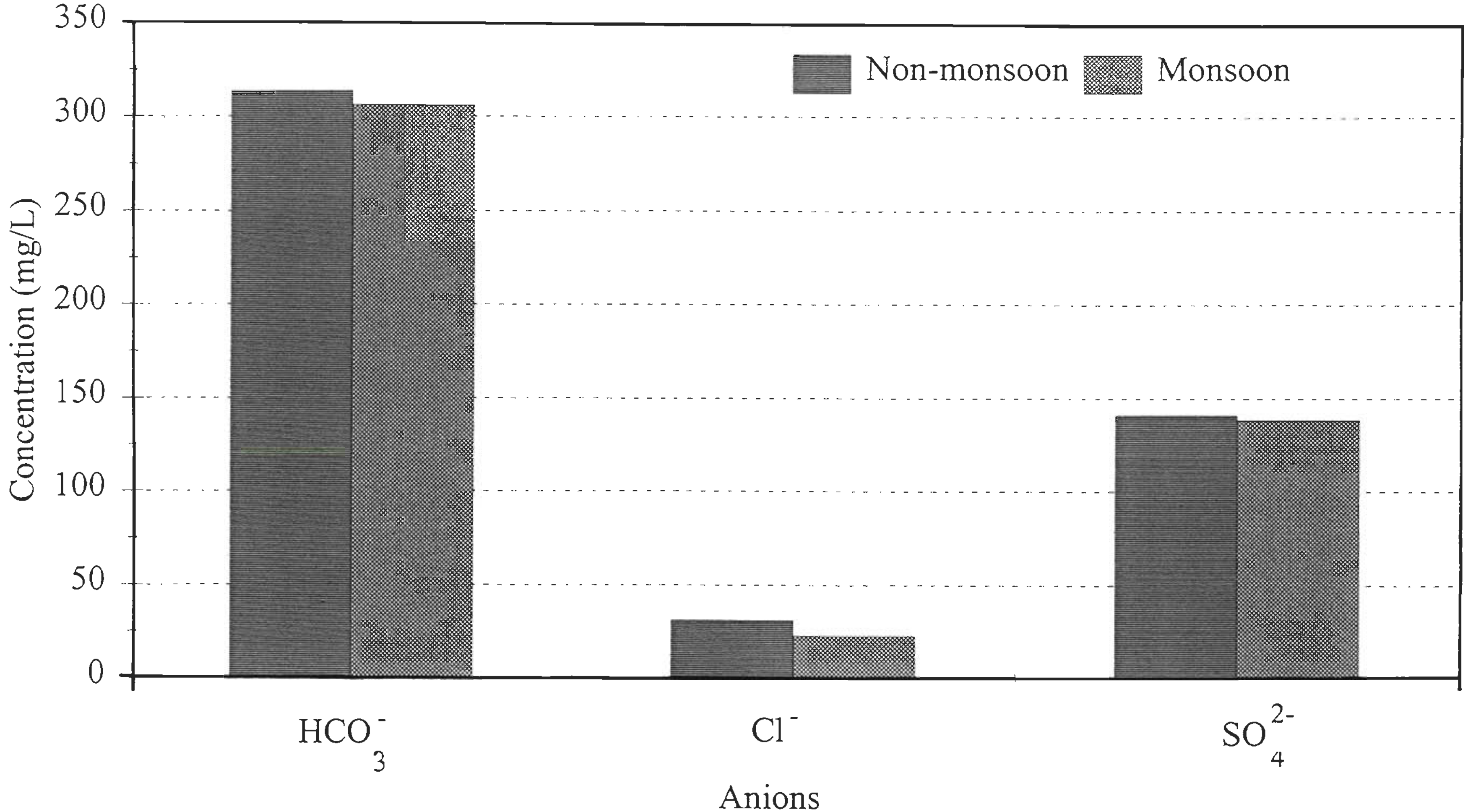


Figure 3.21a Variation in anion concentration in the drains during monsoon and non-monsoon seasons in Nainital lake catchment

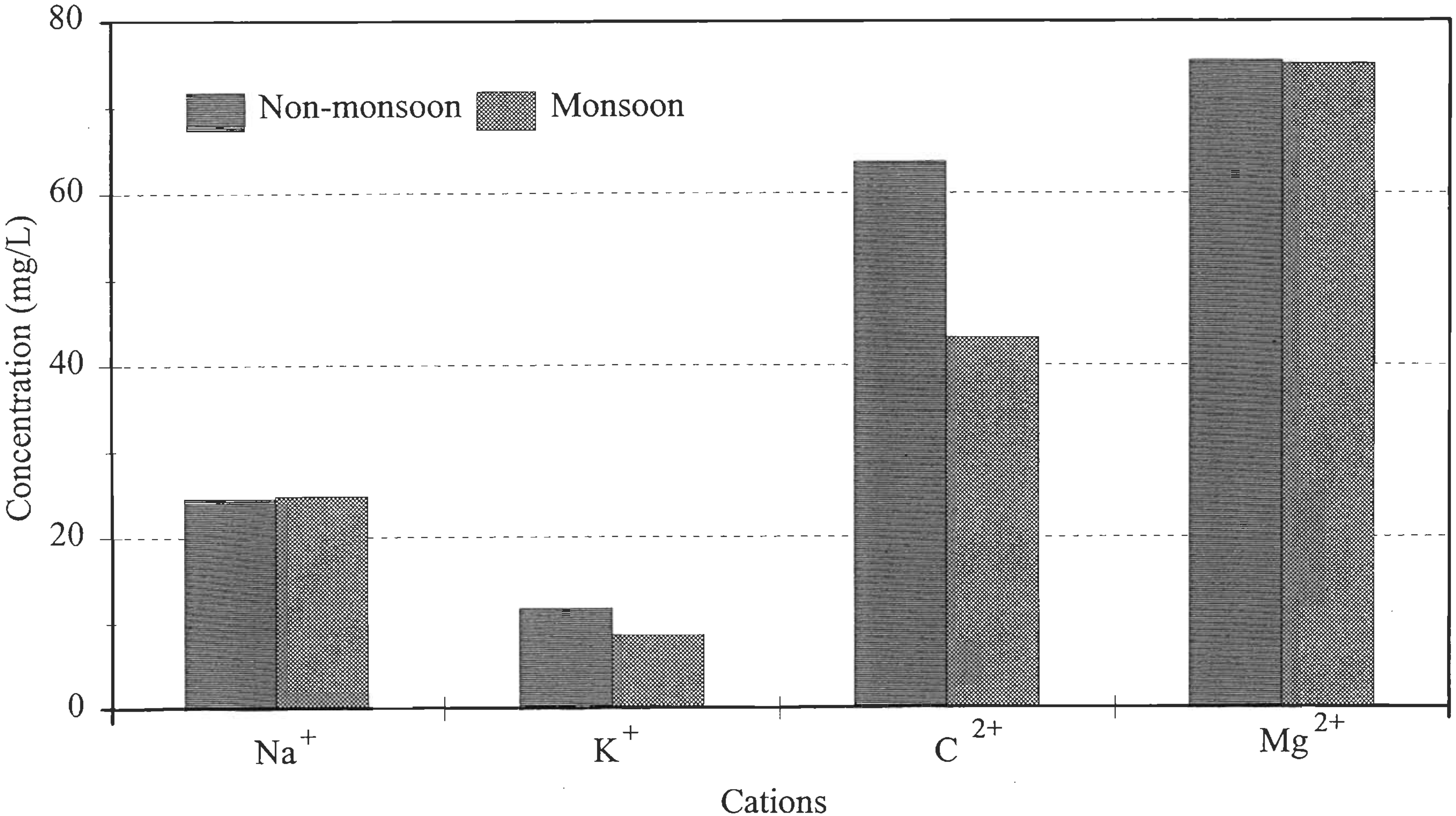


Figure 3.21b Variation in cation concentration in the drains during monsoon and non-monsoon seasons in Nainital lake catchment

and domestic wastes into the drains. The hydrochemical data observed during monsoons may also be considered as representative of the surface inflow as sampling was carried out immediately after rainfall events.

3.4.3 Hydrochemistry of Nainital Lake

To understand the hydrochemistry of Nainital lake, water samples were collected regularly between February 1994 and July, 1996, and analysed for major cations and anions. The number of samples collected from the lake varied between 6 (June, 1995) and 37 (March, 1995). However, keeping in view the need for control points, the points near the deepest portions of Mallital and Tallital sub-basins were sampled regularly during all the sampling periods. Samples were collected from three different depths viz., epilimnion (0.5 m below surface), mesolimnion (6 -9 m below surface) and hypolimnion (1- 4 m above the lake bottom).

In 1994, water samples were collected during four different seasons. However, realising the lake hydrodynamics and the short hydraulic residence time of the lake, samples were collected during 8 months in 1995. In 1996, samples were collected in January, April and July for comparison. The number of samples collected from the lake for analysis during different sampling campaigns are given in Table 3.29.

Table 3.29 Details of samples collected from Mallital and Tallital sub-basins of Nainital lake.

Month	No. of Samples	Month	No. of Samples
Feb.,94	4	Aug.,95	20
May, 94	9	Sep.,95	20
Oct.,94	28	Nov.,95	20
Dec.,94	11	Dec.,95	19
Feb.,95	26	Jan.,96	20
Mar.,95	37	Apr.,96	27
May, 95	18	Jul.,96	8
Jun.,95	6		

Observed range and mean concentration of different anions / cations of the samples collected from the epilimnion and hypolimnion zones of the lake during different seasons viz. winter (December, January & February), pre-monsoon (March, April & May), monsoon (June, July, August & September) and post-monsoon (October & November) are given in Table 3.30. The variation in cation and anion concentrations during different seasons are plotted in Figure 3.22a and Figure 3.22b respectively.

The data show that the lake is of Magnesium Bicarbonate type. It is also seen that the epilimnion and hypolimnion have different concentrations of certain chemical ions during the stratified period. Among the anions, SO_4^{2-} and Cl^- do not show any marked seasonal variation in their concentration in either epilimnion or hypolimnion. Lower concentrations of SO_4^{2-} as compared with surface inflow is probably due to bacterial reduction. The variation in HCO_3^- is due mainly to the phytoplankton activity in the lake. The reduction of HCO_3^- in epilimnion between winter and monsoon seasons is accompanied by an increase in pH. This results in precipitation of Ca^{2+} in the form of $CaCO_3$, which is evidenced by the reduction in the concentration of Ca^{2+} in both epilimnion and hypolimnion. Subsequent increase in the Ca^{2+} in post-monsoon season is then due to increased sub-surface inflow with higher Ca^{2+} concentration. The concentration of Mg^{2+} as well as electrical conductivity (that is a function of total dissolved solids) are also lower during monsoon season as compared to other seasons. This is due to higher inflow to the lake.

Na^+ and K^+ are influenced by anthropogenic activities, unlike Ca^{2+} and Mg^{2+} that are brought into the lake system mainly from dissolution of the limestones / dolomites. Higher concentration of K^+ observed in monsoon as compared to other seasons is mainly due to increased transport of refuse by the storm runoff into the lake. The increase in Na^+ and PO_4-P during both monsoon and post-monsoon seasons also indicate the anthropogenic influence on the lake water quality. It is evident from the above discussion that except for Cl^-, all other ions are not behaving conservatively in the lake environment.

Table 3.30 Hydrochemical data of Nainital lake. The values presented in parentheses indicate the number of samples. All ionic concentrations are in mg/L, Electrical Conductivity (E.C) in μS/cm, and Temperature (T) in °C.

Season *		T	pH	E.C.	HCO_3^-	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+
Winter	Range	7.9 - 11.6	6.4 - 8.7	370 - 680	230 - 350	49 - 145	3 - 32	28.1 - 73.8	32.0 - 96.0	5.4 - 14.4	3.1 - 6.7
	Mean(#)	9.4 (72)	7.5 (51)	500 (71)	275 (52)	93.3 (56)	14 (66)	53.4 (71)	54.6 (71)	9.8 (67)	4.4 (68)
Pre-E	Range	14.9 -21.3	7.8 - 9.4	440 - 890	197 - 268	68 - 126	10 - 26	32.9 - 56.9	44.7 - 63.2	9.4 - 13.6	4.1 - 9.0
	Mean(#)	18.8 (21)	8.5 (14)	600 (21)	237 (18)	89.6 (21)	16 (17)	42.7 (21)	54.5 (21)	11.4 (14)	4.8 (14)
Pre-H	Range	8.2 - 17.2	7.2 - 8.9	360 - 860	234 - 304	64 - 135	6 - 22	36.1 - 80.2	41.3 - 61.2	8.0 - 14.6	4.0 - 5.2
	Mean(#)	9.3 (61)	8.0 (45)	620 (60)	259 (55)	90.7 (61)	15 (55)	49.3 (61)	53.7 (61)	11.5 (45)	4.5 (45)
Mon-E	Range	20.0 - 23.8	6.9 - 9.6	380 - 650	170 - 208	55.4 - 120	0 - 28	21.6 -45.7	31.6 - 53.2	10.8 - 19	3.4 - 7.8
	Mean(#)	21.5 (14)	8.6 (15)	460 (16)	196 (16)	89.1 (14)	14 (16)	33.3 (16)	42.5 (16)	15.1 (14)	5.5 (14)
Mon-H	Range	8.0 - 13.4	7.0 - 9.4	480 - 850	242 - 318	56.4 - 120	0 - 72	26.5 - 64.2	45.7 - 73.9	9.0 - 21.0	3.6 - 8.1
	Mean(#)	10.1 (29)	7.7 (30)	600 (35)	268 (35)	94.1 (29)	21 (34)	39.1 (35)	59.5 (35)	16.0 (29)	6.0 (29)
Post-E	Range	13.1 - 20.5	7.5 - 8.8	510 - 590	174 - 235	74.1- 118	6 - 37	26.5 - 57.7	40.8 - 70.0	13.8 - 39.8	3.5 - 4.9
	Mean(#)	16.4 (13)	8.0 (14)	550 (14)	210 (14)	89.5 (14)	16 (14)	42.6 (14)	52.1 (14)	23.7 (6)	4.1 (6)
Post-H	Range	9.2 - 13.5	6.7 - 8.3	580 - 770	218 - 274	45.3 - 98	5 - 21	42.5 - 80.2	38.4 - 69.0	16.9 - 13.8	4.1 - 5.0
	Mean(#)	10.3 (23)	7.2 (30)	660 (30)	256 (30)	80.2 (25)	13 (30)	59.9 (30)	49.2 (30)	20.6 (10)	4.7 (10)

* Pre - pre-monsoon, mon - monsoon, post - post monsoon E - Epilimnion and H -Hypolimnion.

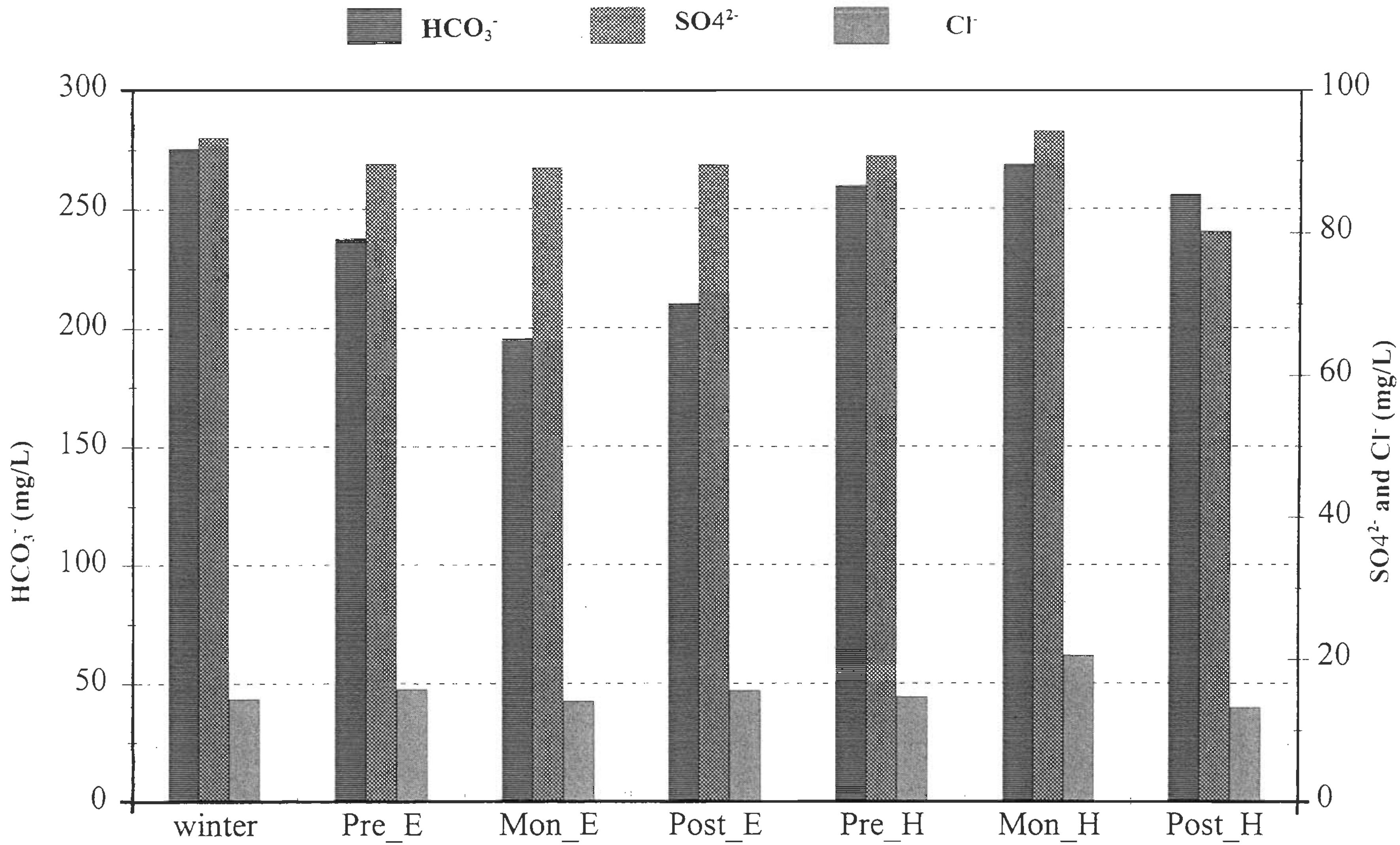


Figure 3.22a Variation in anion concentration in Nainital lake during different seasons.
Pre:pre-monsoon; Post:Post-monsoon; Mon:Monsoon E: Epilimnion and H:Hypolimnion

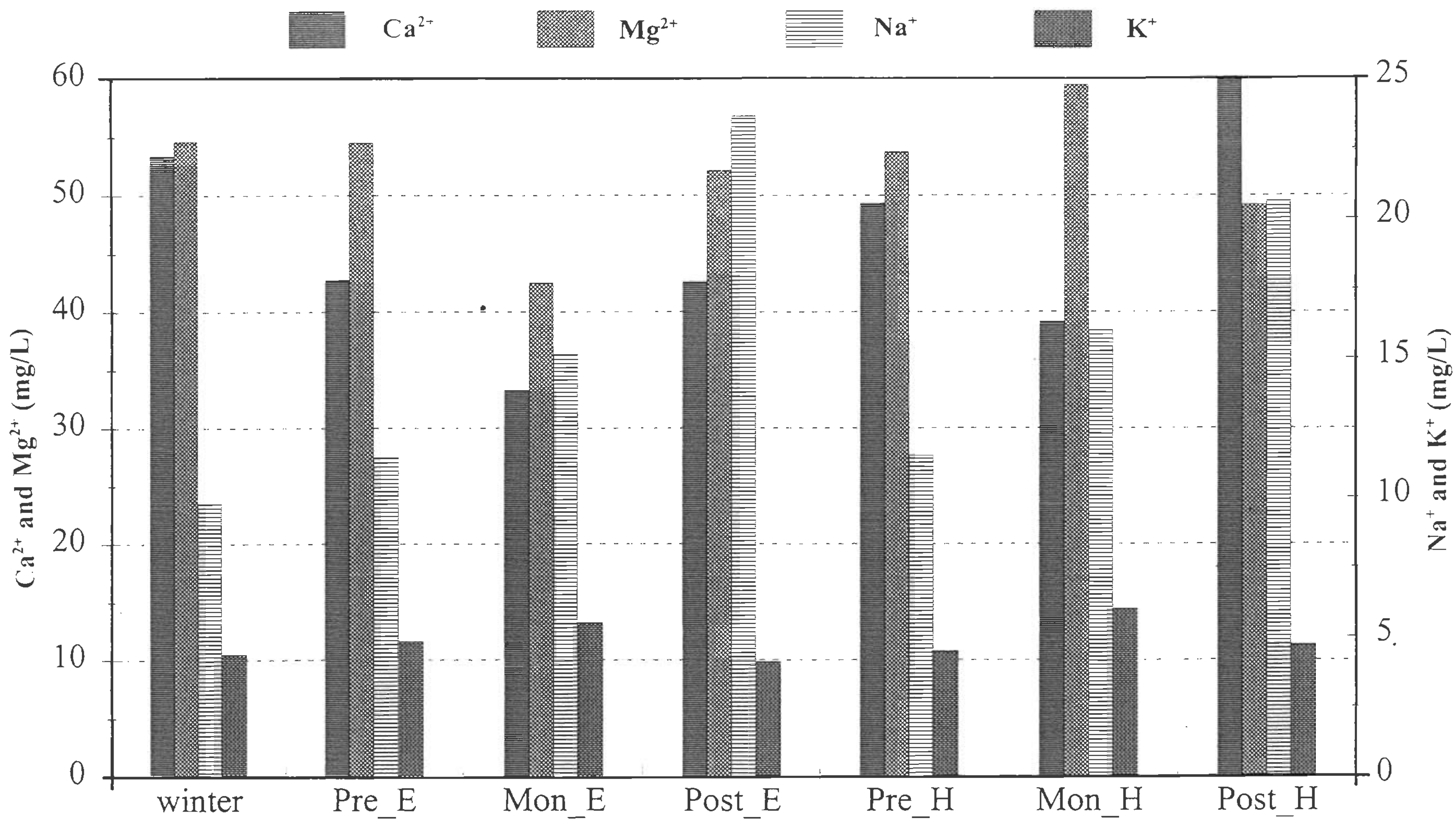


Figure 3.22b Variation in cation concentration in Nainital lake during different seasons.
Pre:pre-monsoon; Post:Post-monsoon; Mon:Monsoon E: Epilimnion and H:Hypolimnion

3.4.4 Hydrochemistry of Groundwater

Groundwater occurring in Nainital lake basin may be grouped into three categories viz., a) upstream springs that are unrelated to Nainital lake, b) the groundwater towards the northern bank of the lake (being tapped by the wells), that is a mixture of groundwater and the lake water, and c) the downstream springs that may or may not be related to the lake. In order to assess the groundwater chemistry and also its relation to the lake, groundwater samples were collected from the upstream and downstream springs. The location of the springs from where the samples were collected are given in Figure 3.23. The hydrochemical data of the upstream springs, wells located in the lake basin and downstream springs are given in Tables 3.31a, 3.31b and 3.31c respectively.

As shales chiefly contain residual minerals cemented usually by calcium carbonates, the principal soluble material in shales are calcium and magnesium carbonates. The chemistry of upstream springs located in shale rocks such as Lakeview, Chunadhara and Doctor House springs is dominated by Ca^{2+} and Mg^{2+}. Doctor House and Chunadhara Springs exhibit relatively larger variations for most of the ions in comparison to Lakeview Spring, which may be due to shorter residence time of groundwater in the former spring region than in the latter spring area.

In carbonate terrains, the equivalent Ca^{2+}/Mg^{2+} ratio (CMR) in groundwater has hydrologic and geochemical significance. In calcite rich rocks, the CMR is generally greater than one, and in dolomite rich rocks the CMR approaches to one (5, 49, 6). However, the CMR for all the springs in the study area, is less than one in all the seasons (Table 3.32). This may be due to the introduction of materials including phosphorus from outside sources or due to incongruent dissolution of dolomite rich rocks with calcite precipitation (93).

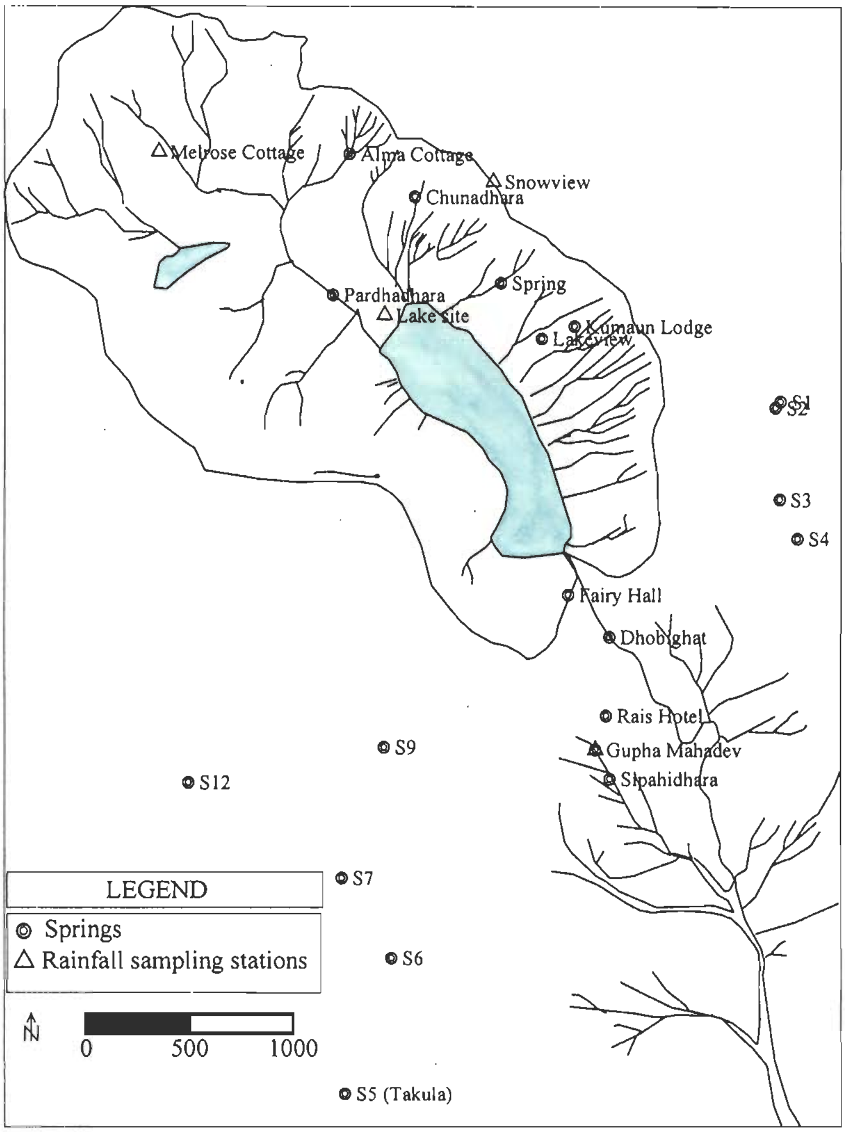


Figure 3.23 Map of Nainital area showing the locations of springs and rainfall sampling stations

Table 3.31a Hydrochemical data of upstream springs. All ionic concentrations are in mg/L, Electrical Conductivity (E.C) in μS/cm

Month	Location	pH	E.C.	HCO_3^-	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+
02/94	PD	-	520	-	-	-	58.0	52.0	-	-
05/94	PD	8.1	845	150	177	6	85.8	64.2	8.6	3.0
12/94	PD	7.7	716	174	-	6	71.0	35.0	13.6	3.6
03/95	LV	-	806	322	184	17	55.1	99.2	10.2	3.1
	PD	-	818	296	176	23	61.1	67.9	12.4	4.5
05/95	LV	7.6	851	266	165	23	28.9	69.0	8.6	2.3
	CD	7.5	943	231	170	21	32.1	67.1	13.0	5.6
	PD	7.3	821	234	170	13	28.9	56.4	13.6	3.3
08/95	LV	-	-	250	100	16	38.5	78.2	14.0	2.6
	DH	-	-	208	86	28	33.7	52.9	12.0	2.2
	PD	-	-	250	144	16	39.3	72.4	18.0	3.7
09/95	LV	-	-	392	184	20	40.9	90.0	13.0	2.9
	CD	-	-	330	150	24	40.1	91.4	22.3	7.8
	PD	-	-	290	134	26	37.7	67.6	14.1	3.7
11/95	LV	8.3	834	332	137	16	49.2	80.0	13.8	2.1
	AC	7.2	788	323	121	30	91.8	68.8	15.9	3.4
	PD	8.0	701	269	122	10	70.2	70.9	20.0	2.9
12/95	LV	8.2	629	380	148	18	81.0	69.0	4.7	2.7
	CD	8.2	625	360	162	22	100.3	60.3	8.0	7.0
01/96	LV	-	-	374	144	-	53.7	94.3	5.2	4.2
	CD	-	-	334	171	18	32.1	100.6	9.6	7.0
	PD	-	-	290	88	12	77.8	64.6	6.7	4.0
07/96	AC	8.1	923	340	-	10	73.0	54.9	-	-

* LV-Lakeview Cottage; CD-Chunadhara; DH-Doctor House; AC-Alma Cottage; PD-Pardhadhara

Table 3.31b Hydrochemical data of groundwater samples from wells located in Nainital lake catchment. All ionic concentrations are in mg/L, Electrical Conductivity (E.C) in μS/cm

Month	Location	pH	E.C.	HCO_3	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+
10/94	OW	7.5	640	238	71	15	62.6	13.1	-	-
12/94	OW	7.7	726	294	-	6	44.0	63.0	14.6	4.4
02/95	OW	6.7	757		73	13	64.9	66.6	12.8	5.0
03/95	SW	-	840	262	168	19	37.9	66.3	9.8	3.5
05/95	OW	7.3	734	262	-	15	36.9	57.8	14.0	4.5
	SW	7.4	764	210	135	17	38.5	49.6	12	3.2
08/95	OW			284	121	46	44.1	74.3	-	-
09/95	OW	-	-	238	102	22	36.9	55.4	21.2	7.5
11/95	OW	7.1	708	237	112	19	70.6	63.9	15.9	3.5
	SW	7.0	679	241	121	9	68.6	68.5	-	-
12/95	OW	8.0	548	364	139	20	72.2	61.2	5.8	3.6
	SW	8.4	530	312	131	12	62.6	58.8	4.7	3.0
01/96	OW	-	-	290	134	14	71.4	61.7	6.8	3.5
	SW	-	-	292	136	14	31.3	71.9	5.2	5.2

* OW-Open well located in Mallital bank; SW-Sukhatal Well located in Sukhatal lake bed.

Carbonate terrains, in which the springs are located, do not have any agricultural or farm activity and therefore, influence of any external source does not exist. However, there is a possibility of sewerage discharge mixing into the groundwater system in the region, which might increase the phosphorous concentration in groundwater. As sewage is also associated with Na^+, CMR was also tested with Na^+. The test showed that there is no significant correlation between CMR and Na^+ and therefore, the presence of external phosphorus appears to be negligible.

Table 3.31c Hydrochemical data of downstream springs located in Balia ravine. All ionic concentrations are in mg/L, Electrical Conductivity (E.C) in μS/cm

Month	Location	pH	E.C.	HCO_3^-	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+
12/94	SD	8.1	560	238	116	6	46.0	48.0	9.2	3.0
	GM	8.3	547	230	119	6	49.0	44.0	8.8	2.7
03/95	SD	-	-	244	150	13	46.4	57.9	9.0	3.8
	GM	-	-	236	133	16	45.9	57.9	10.6	3.0
	RH	-	657	250	117	8	44.7	52.8	14.8	5.5
05/95	SD	7.7	632	230	99	18	36.9	53.5	9.6	3.0
	GM	8.0	602	159	130	11	34.5	38.9	8.0	2.5
06/95	SD	8.2	676	236	100	-	43.3	53.2	8.0	2.6
	GM	7.8	657	230	115	-	50.0	48.6	7.7	1.2
08/95	SD	-	-	218	82	20	36.1	51.5	20.0	4.5
	GM	-	-	228	89	30	36.1	52.0	14.0	2.8
09/95	SD	-	-	236	134	18	36.9	53.9	13.0	3.5
	GM	-	-	220	77	28	32.1	52.9	13.0	3.3
11/95	SD	8.2	582	228	81	10	48.1	60.8	9.7	2.6
	GM	8.6	566	222	83	9	49.3	54.4	10.7	2.3
12/95	SD	8.3	434	288	122	10	48.9	49.1	4.6	3.2
	GM	8.3	468	280	75	16	46.5	56.9	4.3	2.9
01/96	SD	-	-	240	84	18	45.7	64.0	5.3	2.7
	GM	-	-	240	97	14	46.5	58.3	4.7	2.9
06/96	SD	7.9	789	238	-	8	68.2	43.3	-	-
	GM	8.0	796	240	-	10	50.5	51.5	-	-

* SD- Sipahidhara; GM-Gupha Mahadev Temple; RH-Rais Hotel

Comparison of summer and winter 1995 data exhibit an increase in Ca^{2+} by 50%, in Mg^{2+} by 2.7% and in CMR by 51%. The coefficient of variation is lower for Mg^{2+} (7 - 16%) and higher for Ca^{2+} (about 30%) indicating that the variation in CMR is mainly due to the variation in Ca^{2+}. This also suggests that between winters and summers, dolomite undergoes incongruent dissolution with calcite precipitation (93).

Table 3.32 Comparison of average Total Hardness and CMR along with Coefficients of variation for different springs

Location	Total Hardness		Ca^{2+} (epm) / Mg^{2+} (epm)	
	Mean (mg/L)	C.V.(%)	Mean	C.V.(%)
Pardhadhara Spring	406	13	0.49	31
Lakeview Spring	491	13	0.4	38
Chunadhara / Doctor House	386	22	0.33	24
Sariyatal Spring	366	10	0.48	33
Kailakhan Springs (S2/S3/S4)	278	28	0.67	24
Sipahidhara Spring	328	6	0.49	13
Gupha Mahadev Temple Spr.	320	10	0.48	13

Groundwater in near saturation with carbonates may have chemical characteristics, that are related to changes in CO_2 pressure or changes in groundwater chemistry. When CO_2 in groundwater increases, more carbonate rocks dissolve and there is an increase in HCO_3 and decrease in pH. An increase in pH, Ca^{2+} or HCO_3^- cause calcite precipitation with consequent increase in P_{CO2}. This is evident by the negative correlation between HCO_3 and pH (Figure 3.24) and by higher P_{CO2} for the samples collected in summer (-2.45) than in winter (-2.91).

The saturation indices for dolomite (SI_D) and calcite (SI_C) for different springs were calculated from the hydrochemical data by using WATEQ computer program (118), assuming a K_D value of 10^{-17} at 25°C. The results are presented in Table 3.33. Since Nainital groundwater has CMR lower than 0.5, it would be more aggressive (undersaturated) in the dolomite of the study area than in ideal dolomite. Therefore, K_D for dolomitic rocks of Nainital region is calculated by the following equation (71):

$$\frac{(K_C)^2}{K_D} = \frac{[Ca^{2+}]}{[Mg^{2+}]} \qquad (3.22)$$

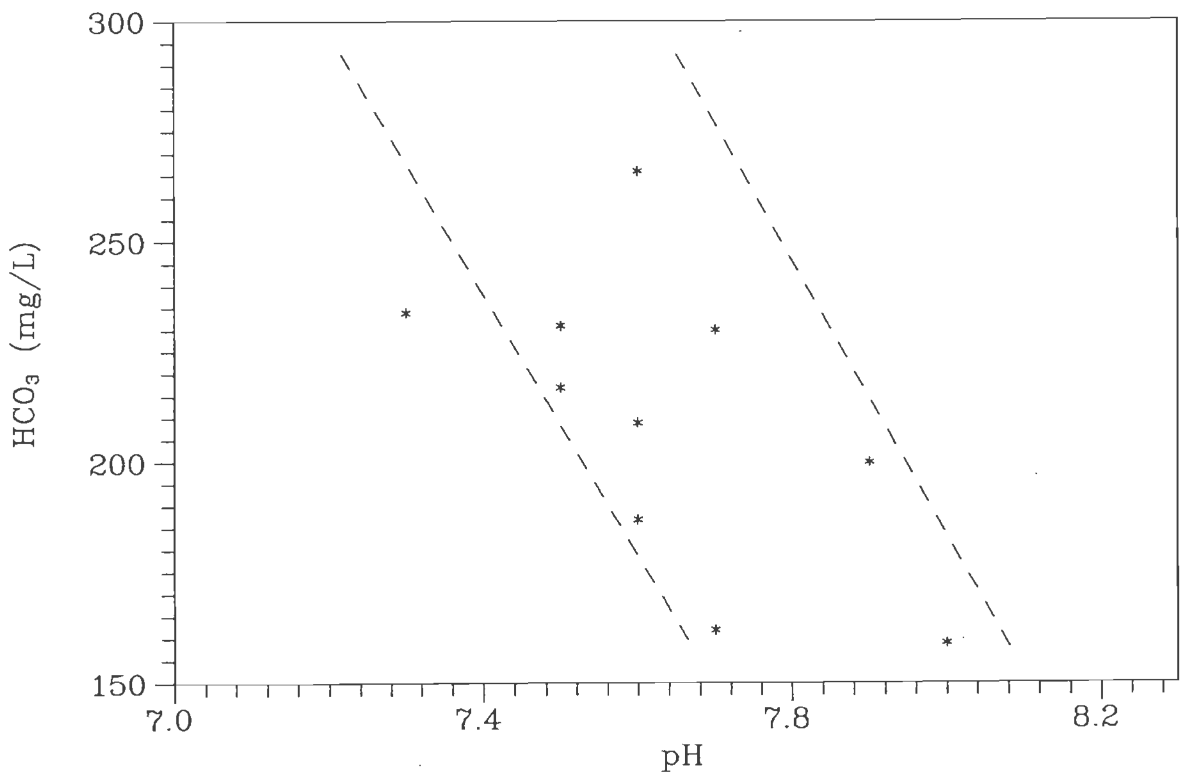


Figure 3.24 pH – HCO_3 cross plot of data pertaining to springs in Nainital area

From the K_C values for different temperatures (93) and a CMR of ~0.5 for Nainital groundwater (15°C), the calculated K_D is $10^{-16.44}$. The CMR at 25°C is ~0.83 and the corresponding K_D value is $10^{-16.72}$. If this K_D value is used in place of K_D for ideal dolomite (16.85 at 15°C), then the SI_D value given in Table 3.33 will be lesser by 0.1. Therefore, the carbonate rocks of Nainital have higher solubility than ideal dolomite.

Table 3.33 Saturation indices of Calcite and Dolomite of spring water sampled during different seasons in 1995.

Season	Spring ID	SI_C	SI_D
Summer, 95	Pardhadhara	-0.55	-0.59
	Chunadhara	-0.33	-0.12
	Doctor House	+0.08	+0.52
	Sipahidhara	-0.03	+0.32
	Gupha Mahadev Temple	+0.09	+0.44
	S3	-0.24	-0.04
	Lakeview	-0.21	+0.17
	Sariyatal	-0.32	-0.10
Winter, 95	Chunadhara	+1.00	+1.98
	Lakeview	+1.12	+2.38
	Sipahidhara	+0.73	+1.67
	Gupha Mahadev Temple	+0.71	+1.71
	Sukhatal well water	+1.32	+2.84

Na^+ (22.3 mg/L), K^+ (7.8 mg/L) and Cl^- (30.0 mg/L) are found in low concentration compared to alkaline earth and sulphate. However, relative variation of Cl^- is very helpful in estimating the evapotranspiration index, ET_{index}. ET_{index} is defined as (98):

$$ET_{Index} = \frac{R_c}{GW_c} \cdot 100 \qquad (3.23)$$

where, R_C and GW_C are concentration of Cl^- in rainwater and groundwater respectively. ET_{index} is used to estimate approximate loss due to evapotranspiration at the time of groundwater recharge. Assuming an average Cl^- concentration of 1.0 mg/L in the Nainital rainfall and using the mean observed Cl^- in groundwater of the Nainital lake catchment (16 mg/L), the average loss due to evapotranspiration is around 94%. In other words, the quantum of water infiltrating into the soil and reaching the zone of saturation is around 6% only.

The presence of sulphate in the groundwater of Nainital lake basin may be attributed to the oxidation of pyrite as noted by earlier workers (101, 26). The oxidation and hydrolysis of pyrite are as follows (78):

$$2FeS_2 + 7H_2O + 15O \rightarrow 2Fe(OH)_3 + 4H_2SO_4$$
$$FeS_2 + H_2O + 7O \rightarrow 2FeSO_4 + H_2SO_4$$
$$2FeSO_4 + H_2SO_4 + O \rightarrow Fe_2(SO_4)_3 + H_2O$$

Sulphuric acid, thus formed, reacts with calcite of the calcareous country rock and produces soluble calcium sulphate and bicarbonate according to the following equation:

$$H_2SO_4 + CaCO_3 \rightarrow H_2CO_3 + CaSO_4$$

Slate/shale and dolomite of Nainital lake basin contain pyrite (26, 8, 152). Therefore, the sulphate in the groundwater is probably derived due to oxidation of pyrites from these rocks.

3.4.5 Chloride Mass Balance

Mass balance can be attempted by means of conservative chemical constituents. Among the various chemical constituents, chloride is a conservative species, and therefore, it may be used for such studies. The advantage of the chloride mass balance over the isotope mass balance method, is that the mass of chloride loss from the lake through evaporation is zero. Therefore, it is much simpler as compared to isotope mass balance. However, chloride may be introduced into the lake and groundwater systems through anthropogenic activities and becomes disadvantageous. In the present work chloride mass balance has been attempted for the purpose of comparison.

The concentration of chloride in lake water was 8 mg/L and 10 mg/L during February 94 and 95 respectively. The mean concentration in drain water (D_I) was 31 mg/L, and that in surface inflow (S_I) was 24 mg/L. The latter value has been used considering the fact that during monsoon period the flow in the drain water, particularly during the sampling period, was dominated by channelled surface runoff. The mean concentration of chloride in groundwater was 16 mg/L in the upstream springs viz., Pardhadhara, Alma cottage and Lakeview springs. The mean chloride concentration (17 mg/L) of the downstream springs, Sipahidhara and Gupha Mahadev Temple, has been considered as representative of subsurface outflow from the lake. The input of chloride by precipitation has been considered as 1 mg/L (cf. Section 3.4.1).

Sub-surface outflow (SS_O) from the lake, computed using the chloride mass balance approach, has been presented in Table 3.41. The result has been used in Equation (3.32) to compute sub-surface inflow (SS_I) to the lake. The results corroborate the findings of conventional water balance method i.e., the sub-surface components are dominant over other components. The SS_I and SS_O computed using the chloride mass balance method account for about 55.0% of total inflow and about 59.0% of total outflow respectively. As compared with the estimates of conventional water balance, the SS_O computed by chloride mass balance method is higher by 5%.

3.5 Isotope Mass Balance

3.5.1 Stable Isotope Ratios of Precipitation

In order to understand the stable isotope characteristics of precipitation in Nainital area, rainfall samples were collected from four stations set up at different altitudes. Among these three were within the basin and one on the downstream side (Figure 3.23).

3.5.1.1 Local meteoric water line

The statistical analysis of rainfall isotopic data (Table 3.34) for Nainital area yielded the following equation for monsoon period:

$$\delta D\ ‰ = 7.5 \cdot \delta^{18}O + 4.82 \quad (n = 15;\ r = 0.97) \qquad (3.24)$$

This equation compares well with that proposed by other workers (16, 130, 85, 12). Based on large number of groundwater samples from northern India, a regional meteoric water line with a slope of ~7.2 was proposed as a characteristic of Indian monsoon (85). An equation with a slope of ~7.1 for Gaula basin of the Kumaun Himalaya, has been developed based on randomly selected individual storm samples instead of monthly-integrated rainfall samples (12).

Compared to Equation 2.5 of the GMWL, the differences in slope and intercept of equation 3.24 are due to secondary evaporation prevalent in Indian sub-continent (16, 85, 36).

Table 3.34 Isotopic composition of rainfall samples collected at different altitudes in the Nainital area.

Site	Elevation, m	Month	$\delta^{18}O$, ‰	δD, ‰	'd', ‰
Snowview	2275	July, 1995	-12.3	-87.0	11
		Aug, 1995	-12.4	-90.6	9
		Sep, 1995	-12.2	-83.6	14
Mel Rose Cottage	2140	July, 1995	-11.9	-83.8	11
		Aug, 1995	-11.8	-87.4	7
		Sep, 1995	-11.3	-80.0	10
Lake site	1940	July, 1995	-11.4	-79.2	12
		Aug, 1995	-11.3	-83.2	7
		Sep, 1995	-10.5	-77.1	7
Lake Site*	1940	Sep., 1994	-12.6	-88.1	13
		Apr., 1995	-1.6	-6.8	6
		May, 1995	-10.5	-64.1	20
Gupha Mahadev Temple	1830	July, 1995	-10.7	-76.6	9
		Aug, 1995	-11.0	-78.0	10
		Sep, 1995	-9.5	-70.3	5

* This set of data has not been used to study the altitude effect, but used in defining LMWL.

3.5.1.2 *'d' - excess parameter*

The isotopic composition of source area controls the 'd' excess at cloud base. Therefore, the 'd'-excess values of precipitation samples collected at different elevations should be comparable to each other. In other words, the 'd'-excess data should exhibit poor correlation with altitude - if secondary evaporation is not significant. A perusal of Table 3.34 indicates that this is found to be true for July - August precipitation, but not so for the month of September. The altitude effect in rainfall isotopic composition, calculated for monthly rainfall isotope data, are presented in Table 3.35. The correlation coefficient (r) of isotope values and 'd' excess versus the corresponding altitude are also given in Table 3.35. The September 1995 data exemplifies the 'secondary-altitude effect' caused by secondary evaporation (25).

Table 3.35 Estimated altitude effects (per 100 m) in $\delta^{18}O$ and δD for the monsoon precipitation of 1995 at Nainital, Kumaun Himalaya.

Month	$\delta^{18}O$, ‰	δD, ‰	'd' excess, ‰
July	0.34 ± 0.05 r=0.96	2.3 ± 0.01 r=0.99	- r=0.33
August	0.31 ± 0.03 r=0.98	2.7 ± 0.35 r=0.97	- r=0.14
September	0.57 ± 0.06 r=0.98	2.7 ± 0.57 r=0.92	- r=0.98

3.5.1.3 Altitude effect

The isotopic data of 1995 monsoon season presented in Table 3.34 are used for the estimation of mean altitude effect. The elevation versus isotopic ratios are plotted in Figures 3.25a and 3.25b. The estimated altitude effect for different months in $\delta^{18}O$ and in δD are presented in Table 3.35. It is observed that the $\delta^{18}O$ data shows a higher slope for September than for July or August. As discussed earlier, this higher slope is probably due to the effect of secondary evaporation of falling raindrops and the evaporation effect appears to be linearly related to the distance travelled by the droplets through air (Table 3.35). On the other hand, the δD data of September do not show this effect, because for δD the kinetic fractionation factor plays a minor role, while in case of $\delta^{18}O$ the equilibrium and kinetic fractionation factors have the same magnitude. From the above discussion, it is evident that two different processes cause the altitude effect in the study area. Therefore, the mean altitude effect in the rainfall isotope data

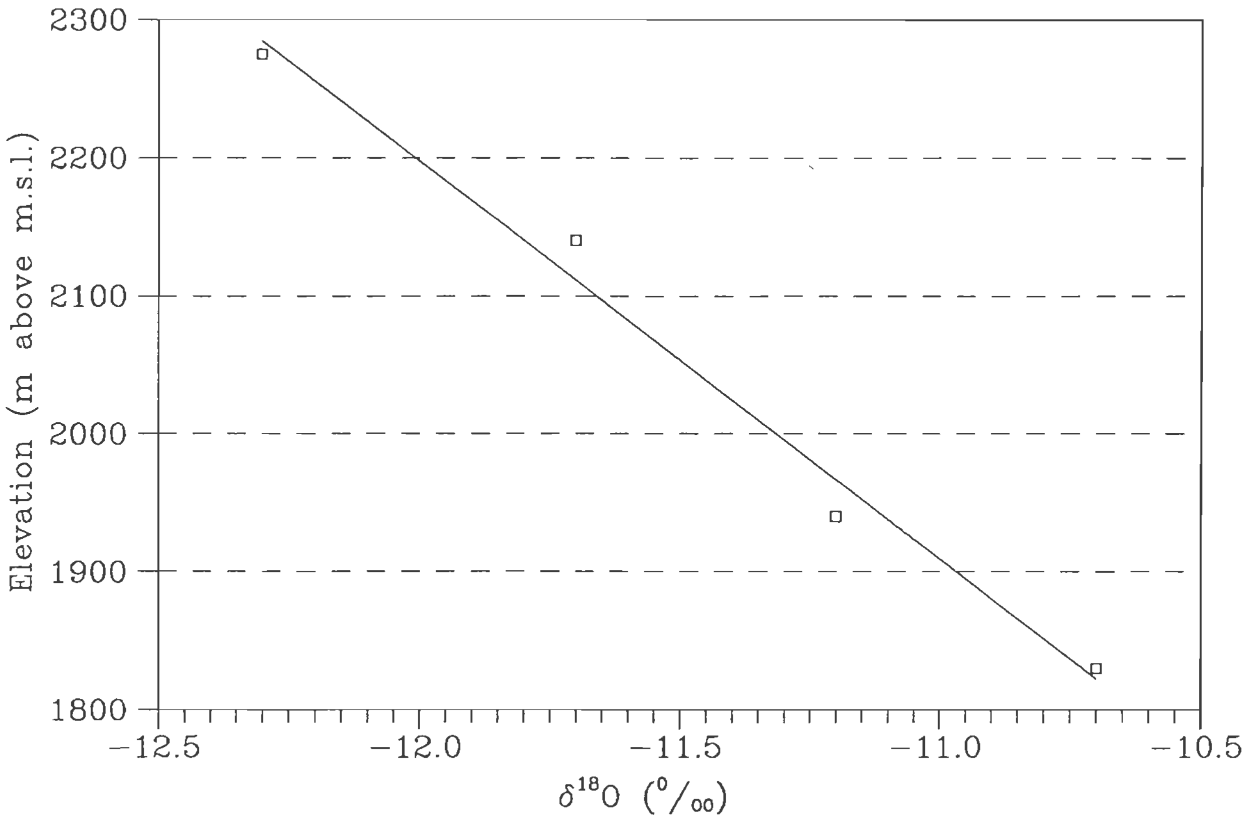


Figure 3.25a Altitude effect in $\delta^{18}O$ during the monsoon season of 1995 based on rainfall weighted averages

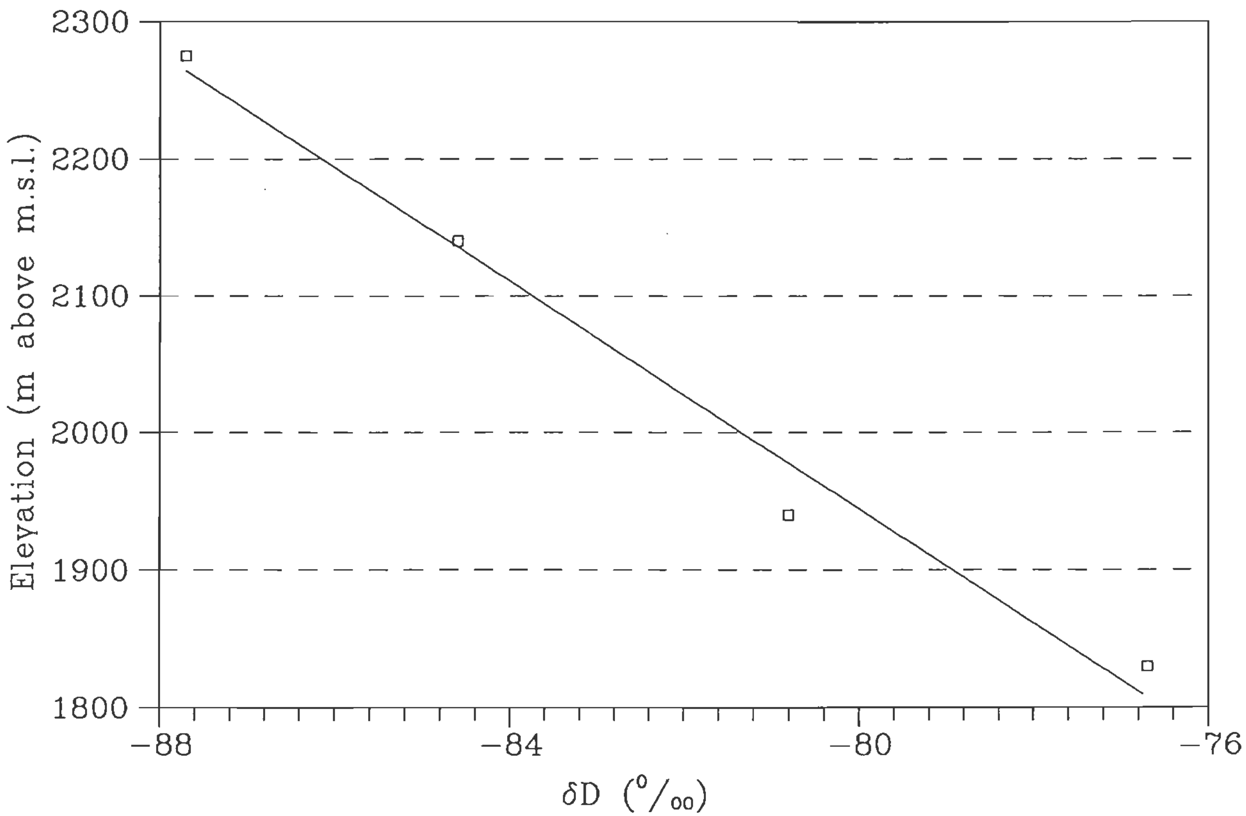


Figure 3.25b Altitude effect in δD during the monsoon season of 1995 based on rainfall weighted averages.

has been estimated by the rainfall weighted average method that yielded a value of -0.34 ± 0.10‰ per 100 m [r=0.99] for $\delta^{18}O$ and -2.4‰ ± 0.8‰ per 100 m [r=0.98] for δD.

The mean altitude effect presented above is within the range of values reported from other parts of the world (162, 23). However, the values are higher than altitude effect of -0.19‰ for $\delta^{18}O$ and -1.6‰ for δD, reported by other researchers based on the samples from river Ganga (120). These lower values can be ascribed to the contribution from groundwater that is recharged at different altitudes and also evaporation of river water. The mean altitude effect in $\delta^{18}O$, observed in the present study is also higher than the altitude effect estimated by using samples from springs in the Gaula basin (12).

3.5.1.4 Local meteorological conditions and δp

Isotopic ratios of precipitation, if not available for a particular location/period, could be generated using the multiple linear regression equations using IAEA/WMO network data (162). However, to achieve this, regional scale climatological parameters are required for getting better results (162).

In the present study, isotopic ratios of precipitation are generated for the months for which the observed data were not available. For this purpose, a multiple regression equation was developed using monthly mean values of temperature and relative humidity, monthly rainfall and isotopic data observed at the lake site (Table 3.36). The equation yielded good results with a multiple correlation coefficient of 0.88 and a standard error of 2.0‰. The equation is as follows:

$$\delta^{18}O\ ‰ = (-0.3)*T_m+(-0.27)*RH_m + (0.01)*RF + 8.98 \qquad (3.25)$$

$$(n=7;\ R=0.88)$$

where, T_m = mean monthly temperature (°C),

RH_m = mean monthly relative humidity (%) and

RF = monthly rainfall (mm).

Table 3.36 Isotopic composition of rainfall / snow and the monthly meteorological data used for developing the multiple regression equation.

Month	$\delta^{18}O$ (‰)	T_m (°C)	Rh_m (%)	RF (mm)
September, 1994	-12.6	17.7	68.4	43
February, 1995	-7.6	6.4	58.6	59
April, 1995	-1.6	15.3	32.6	6
May, 1995	-10.5	20.2	40.9	19
July, 1995	-11.4	18.8	82.3	509
August, 1995	-11.3	18.4	85.5	632
September, 1995	-10.5	17.2	77.9	341

3.5.2 Stable Isotope Ratios of Drain Water

A total of eleven samples were collected from the drains during the study period for stable isotope analysis. Out of these four represent the perennial Nainadevi drain and another four represent the Rickshaw Stand drain. The remaining three samples represent the seasonal drains viz., Boatclub drain and Library drain. The results are given in Table 3.37.

Table 3.37 Isotopic characteristics of drain water

Drain ID	$\delta^{18}O$ (‰)			δD (‰)		
	#	Mean	σ	#	Mean	σ
Nainadevi/ R. Stand	4	-8.6	0.6	4	-52	7
Boat Club/ Library	3	-8	0.6	2	-57	-

It would be seen that the observed value of $\delta^{18}O$ ranges from -7.4‰ to -9.6‰, while that of δD from -44‰ to -64‰. If we consider all the drains, then the average values of $\delta^{18}O$ and δD are -8.4‰ (σ=0.7) and -54‰ (σ=7), respectively. The measured $\delta^{18}O$ and δD values of the drains are considerably higher than those of the local precipitation indicating that some amount of evaporative enrichment might have taken place. The slope of the $\delta^{18}O$ - δD line is ~5.4, which is much less than that of LMWL (Equation 3.24). This also indicates that the drain water might have suffered evaporative enrichment.

3.5.3 Stable Isotope Ratios of Lake

Water samples for stable isotope analysis were collected from the lake during different months of the study period. The results are given in Table 3.38.

The equation for the best-fit line using the $\delta^{18}O$ and δD data pertaining to the lake is as follows:

$$\delta D = (7.1 \pm 0.4) \cdot \delta^{18}O + (2.3 \pm 2.6) \quad (n = 131;\ r = 0.74) \tag{3.26}$$

The above equation is very close to that of the LMWL (Figure 3.26) indicating that the lake has not been significantly affected by non-equilibrium evaporative enrichment processes. The effect of evaporative enrichment is confined mainly to the epilimnion zone. The annual mixing of hypolimnion water and also groundwater, both of which are less affected by evaporation, has probably contributed to the above relation between $\delta^{18}O$ and δD.

Table 3.38 Mean $\delta^{18}O$ and δD data of epilimnion and hypolimnion zones of Nainital lake during different months.

Month of sampling	$\delta^{18}O$		δD	
	Epilimnion $x \pm \sigma$ (#)	Hypolimnion $x \pm \sigma$ (#)	Epilimnion $x \pm \sigma$ (#)	Hypolimnion $x \pm \sigma$ (#)
February,94	-8.2 ± 0.4 (2)	-8.1 ± 0.1 (2)	-49 ± 6 (2)	-55 ± 2 (2)
May, 94	-6.2 ± 0.4	-7.2 ± 0.9	-	-
October, 94	-5.9	-7.3	-	-
December,94	-9.7 ± 0.1 (7)	-9.8 ± 0.1 (2)	-	-
February,95	-7.2 ± 0.4 (3)	-7.4 ± 0.3 (22)	-48 ± 2 (3)	-52 ± 3 (22)
March, 95	-6.0 ± 0.4 (6)	-7.2 ± 0.5 (20)	-42 ± 1 (6)	-49 ± 4 (20)
May, 95	-6.2 ± 0.5 (15)	-7.3 ± 0.9 (22)	-39 ± 5(15)	-48 ± 9 (22)
June, 95	-5.6 ± 0.3 (6)	-7.1 ± 0.7 (11)	-35 ± 1 (6)	-46 ± 6 (11)
August, 95	-5.5 ± 0.3 (9)	-6.8 ± 1.1 (16)	-38 ± 4 (9)	-47 ± 8 (16)
November,95	-8.4 (3)	-8.0 (1)	-	-
April, 96	-7.0 ± 0.4 (2)	-7.7 ± 0.5 (4)	-	-
September, 96	-8.1	-7.6 (2)	-53	-50 (2)

* The values given in parentheses represent the number of samples.

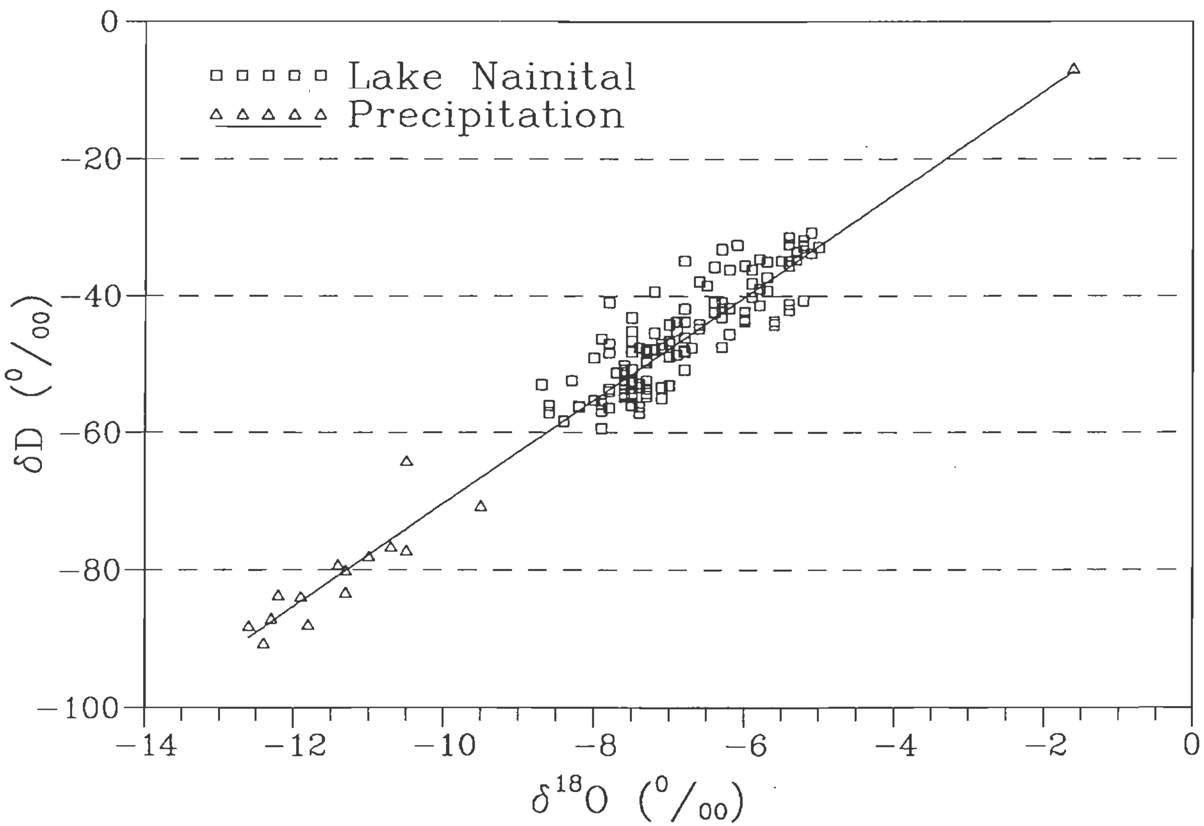


Figure 3.26 Stable isotopic composition of Nainital lake alongwith LMWL.

3.5.4 Stable Isotope Ratios of Lake Evaporates

The stable isotope composition of lake evaporate (δ_E) is not directly measurable, but can be calculated using the Craig and Gordon Linear Resistance (CGLR) Model. The simplified evaporation model (28) can be written as:

$$\delta_E = \frac{(\alpha^* \delta_L - h\delta_a - \epsilon)}{(1 - h + 10^{-3}\Delta\epsilon)} \tag{3.27}$$

where, δ_L is the isotopic composition of lake surface, δ_a is the isotopic composition of atmospheric water vapour, h is the relative humidity normalised to temperature at the lake water surface, α^* is the equilibrium fractionation factor, and ϵ & $\Delta\epsilon$ are enrichment factors.

In the present study, the fractionation and enrichment factors have been calculated by standard methods (95, 63) from the observed mean monthly temperature data. The isotope value observed in the epilimnion zone of the lake has been considered as δ_L, while δ_a has been estimated by the δ_a-δ_p Equilibrium Assumption method (166, 167):

$$\delta_a = \alpha^* \delta_p - \epsilon^* \tag{3.28}$$

where, δ_p is the isotopic composition of local precipitation. For the period for which actual δ_p values were not available, δ_p has been calculated using the multiple linear equation (3.25). Calculated δ_E values are presented in Table 3.39. δ_E values for the months of January to April, 1995 and December, 1995 have not been calculated using the CGLR model as the limiting condition is not satisfied (87).

Table 3.39 Values of different input parameters used in the CGLR Model alongwith calculated values of δ_E for the lake Nainital.

Period	δ_L (‰)	δ_P (‰)	δ_a (‰)	h (%)	α	ϵ^*	$\Delta\epsilon$	δ_E (‰)
01/95	-	-5.6	-16.7	57.8	0.9888	11.13	6.34	-
02/95	-8.4	-7.6	-18.4	58.6	0.9891	10.93	6.08	-
03/95	-7.1	-4.9	-15.4	47.0	0.9894	10.58	7.69	-
04/95	-	-1.6	-11.7	32.6	0.9899	10.10	9.67	-
05/95	-6.2	-10.5	-20.1	40.9	0.9903	9.68	8.44	-26.8
06/95	-5.5	-12.6	-22.0	61.4	0.9904	9.54	5.57	-17.9
07/95	-5.7	-11.4	-21.1	82.3	0.9902	9.80	2.78	-6.4
08/95	-	-11.3	-21.0	85.5	0.9902	9.83	2.31	-11.9
09/95	-8.1	-10.5	-20.3	77.9	0.9901	9.94	3.42	-24.3
10/95	-	-10.4	-20.4	58.1	0.9899	10.10	6.19	-29.7
11/95	-8.2	-8.4	-18.7	54.9	0.9896	10.38	6.60	-32.1
12/95	-8.4	-8.0	-18.6	59.4	0.9893	10.71	5.98	-

From the results presented in Table 3.39, it is seen that the δ_E values are controlled mainly by the relative humidity.

3.5.5 Stable Isotope Ratios of Groundwater

The $\delta^{18}O$ analysis was carried out on samples collected from upstream springs and also from the wells located on the lake bank. The results are given in Table 3.40. It is seen that the springs show considerable variation in $\delta^{18}O$ during different seasons. The variation in $\delta^{18}O$ is larger for the Pardhadhara spring (-7.3‰ to -8.8‰). This is probably due to the presence of Sukhatal lake, the temporary lake in the recharge area of Pardhadhara spring. Sukhatal lake water, which is isotopically enriched due to evaporation during June to September, probably got mixed with the sub-surface reservoir. The $\delta^{18}O$ of the Pardhadhara spring shows a depleted value in winter 1994 than in post-monsoon 1995, probably due to the infiltration of the snowmelt water. However, this could not be verified due to non-availability of the isotopic composition of 1994 snowfall.

Table 3.40 $\delta^{18}O$ data of upstream springs during different seasons.

Season	Month	Spring ID	$\delta^{18}O$ ‰
Winter	February, 94	Pardhadhara	-8.2
Pre-Monsoon	March, 95	Pardhadhara	-7.5
	May, 95	Pardhadhara	-7.7
	April, 96	Sariyatal	-7.4
Monsoon	June, 95	Lakeview	-8
	June, 95	Doctor House	-9.3
	August, 95	Lakeview	-9
	September, 95	Doctor House	-9.4
	September, 95	Lakeview	-8.3
	September, 96	Pardhadhara	-8.8
	September, 96	Alma Cottage	-9.1
	September, 96	Chunadhara	-9.1
Post-Monsoon	November, 95	Lakeview	-8.3
	November, 95	Alma Cottage	-8.2
	November, 95	Pardhadhara	-7.3

It is also seen that Chunadhara, Alma Cottage and Doctor House springs have $\delta^{18}O$ values of -9.1 to -9.4‰ during monsoons. Compared to these the $\delta^{18}O$ values of the Lakeview spring varies between -8.0‰ and -9.0‰ indicating that the spring has more than one source. The post-monsoon $\delta^{18}O$ values of Alma Cottage and Lakeview springs are comparable (-8.3‰) and are heavier than $\delta^{18}O$ value of Alma Cottage spring during monsoons. It indicates delayed recharge of groundwater, which has undergone enrichment due to partial evaporation. The frequency distribution analysis of $\delta^{18}O$ show that for monsoon and non-monsoon seasons the peak values are -9.0‰ and -8.2‰ respectively indicating that the non-monsoon recharge results in isotopic enrichment of groundwater.

Compared to the weighted mean $\delta^{18}O$ of the monsoon season (-11.3‰) the monsoon groundwater mean $\delta^{18}O$ (-9.0‰) shows an enrichment of 2.3‰. Further, the use of multiple regression equation (3.25) with the data pertaining to the months that had a rainfall in excess of 10 mm, yields a $\delta^{18}O$ value of -9.0‰ comparable to the mean observed $\delta^{18}O$ of the groundwater.

Since infiltration rates in the catchment area are quite variable, it is possible that the rain falling in zones of higher infiltration capacity percolates and reaches the groundwater regime more quickly than the rain falling in zones of lower infiltration capacity. Although the distribution of zones of different rate of infiltration is not well defined in the catchment area, the infiltration is higher in linear depressions coinciding with natural drains. The surfacial sheet flow over less permeable soil cover and the interflow may be heavier in $\delta^{18}O$ than the water that percolates into the groundwater regime. However, both these pathways lead to an overall $\delta^{18}O$ enrichment of the groundwater.

The isotopic enrichment in groundwater may also be due to the flushing of soil water, which has undergone evaporative enrichment by the infiltration of subsequent rainfall. This effect is more pronounced on barren top soil than the top soil covered with grass (166). Since approximately 50% of the catchment area of Nainital lake is characterised by non-forestry land use, it is possible that the soil water suffers enrichment due to evaporation. Further, it is also not uncommon that a difference of about 2.0‰ is observed between the $\delta^{18}O$ of rainfall and that of the local groundwater, as similar differences have been reported in other areas (88, 91).

3.5.6 Isotope Mass Balance

Isotope mass balance of a lake may be written as

$$\Delta\delta V = (\delta_P P_I + \delta_{Si} S_I + \delta_{Di} D_I + \delta_g SS_I) - (\delta_E E_O + \delta_{So} S_O + \delta_L SS_O) \qquad (3.29)$$

Equation 3.29 can be rearranged to get sub-surface terms:

$$\delta_g SS_I - \delta_L SS_O = (\delta_E E_O + \delta_{So} S_O \pm \Delta\delta_L V) - (\delta_P P_I + \delta_{Si} S_I + \delta_{Di} D_I) \qquad (3.30)$$

where, SS_I, SS_O, E_O, S_O, P_I, S_I, D_I and ΔV are as given in Equation (3.20) and δ_g, δ_{Go}, δ_E, δ_{So}, δ_P, δ_{Si}, δ_{Di} and δ_L are the corresponding isotopic values. Rearranging Equation (3.20) and solving simultaneously with Equation (3.30), we get:

$$SS_O = \frac{[\delta_G (S_O + E_O \pm \Delta V - S_I - D_I - P_I) - (\delta_L S_O + \delta_E E_O \pm \Delta \delta_L V - \delta_P P_I - \delta_D D_I - \delta_S S_I)]}{(\delta_L - \delta_G)} \qquad (3.31)$$

The above equation is used to determine the sub-surface outflow component of the lake, which in turn is used to estimate groundwater inflow to the lake by the following relation:

$$SS_I = [(E_O + S_O \pm \Delta V) - (P_I + D_I + S_I)] + SS_O \qquad (3.32)$$

This method of estimation does not require prior estimation of the outflow from the lake through springs and pumping wells. Isotope mass balance has been attempted for the period between February, 1994 and February 1995. Since in the month of February, the lake reamins well mixed and homogeneous, it eliminates the stratification effects on the calculation. The mean $\delta^{18}O$ values of the lake considered for mass balance are –8.2‰ (February, 1994) and –7.3‰ (February, 1995) with a net change of 0.9‰. The $\delta^{18}O$ values for different components are precipitation -11.3‰, evaporation -29.1‰, surface inflow -8.6‰ and inflow through the drains -8.0‰. The $\delta^{18}O$ of surface outflow is taken as -8.0‰ as surface outflow occurs mostly at higher lake water levels and with higher surface inflow, having less time for proper mixing. This is shown by the values observed during September 1996, when the surface layers were comparatively depleted than the bottom water. The $\delta^{18}O$ value of groundwater inflow is -9.0‰ and that of the subsurface outflow from the lake is -8.0‰.

Sub-surface outflow (SS_O) of the lake calculated by isotopic mass balance method is presented in Table 3.41. The results are used in equation (3.31) to compute groundwater inflow (SS_I) to the lake. The results indicate that sub-surface components are dominant over other components. The SS_I and SS_O account for 51% and 56% of total inflow and total outflow respectively.

3.5.7 Uncertainties in Isotopic Mass Balance Approach

Uncertainty in the estimation of sub-surface components of the lake can be estimated by using a general expression (15):

$$\sigma_x^2 = \sigma_u^2 \left(\frac{\partial x}{\partial u} \right)^2 + \sigma_v^2 \left(\frac{\partial x}{\partial v} \right)^2 + \ldots\ldots\ldots$$

Since SS_O is a function of all inflow (except SS_I) and outflow components including $\delta^{18}O$ data, the standard error in all these parameters should be known. The standard errors in water balance components have been presented in Table 3.24. The standard error in the $\delta^{18}O$ values of the components are: lake 0.6‰, precipitation 0.2‰, surface inflow 0.6‰, inflow through drains 0.6‰, and surface outflow 0.6‰. The error term associated with the $\delta^{18}O$ values of lake evaporates, cannot easily be ascertained.

In the estimation of δ_E (Equation 3.27), three variables are used viz., δ_l, δ_a and h. The error involved in the determination of these variables will be propagated in the estimated isotopic composition of lake evaporate. In the case of a well-mixed lake, spatial variation of δ_l may not be significant. The error in the computation of δ_a may be as large as 2.5‰ (87). However, error in the measurement of third variable h, and its mean value is considerable (94). The propagated error in the estimation of δ_E, due to the errors (σ^2) in the input variables can be described as (15)

$$\sigma_{\delta_E}^2 = \sigma_{\delta_L}^2 \left(\frac{\partial \delta_E}{\partial \delta_L} \right)^2 + \sigma_{\delta_a}^2 \left(\frac{\partial \delta_E}{\partial \delta_a} \right)^2 + \sigma_h^2 \left(\frac{\partial \delta_E}{\partial_h} \right)^2 \qquad (3.33)$$

From equation (3.27), we get

$$\frac{\partial \delta_E}{\partial \delta_L} = \frac{\alpha^{\cdot}}{(1 - h)} \qquad (3.34)$$

$$\frac{\partial \delta_E}{\partial \delta_a} = \frac{-h}{(1 - h)} \tag{3.35}$$

$$\frac{\partial \delta_E}{\partial h} = \frac{(\delta_L - \delta_a - \epsilon')}{(1 - h)^2} \tag{3.36}$$

Combining equations (3.34), (3.35) and (3.36),

$$\sigma^2_{\delta_E} = \sigma^2_{\delta_L} \frac{\alpha'}{1 - h} + \sigma^2_{\delta_a} \frac{-h}{(1 - h)^2} + \sigma^2_h \frac{(\delta_L - \delta_a - \epsilon')}{(1 - h)^2} \tag{3.37}$$

From the first two terms on the right hand side of equation (3.37), it is seen that the propagation of errors in δ_L and δ_a are controlled by the values of humidity. When h approaches unity, the error becomes infinite. The propagated error in δ_E due to the error in δ_a will be slightly higher than that in δ_L, for the same input error as seen in Figure 3.27. The propagated errors due to errors in input isotopic data, δ_L and δ_a, will result in unrealistic values of δ_E, only when the humidity exceeds 0.8. On the other hand, small error in input variable h will result in unrealistic values of δ_E when the humidity exceeds 0.75. The estimated error for δ_E is about 4‰. The error in the estimated sub-surface outflow from the lake, computed by using the standard errors in the input variables, is large indicating the sensitivity of the approach.

The isotope mass balance method is sensitive to the difference between the $\delta^{18}O$ values of groundwater inflow and that of the lake seepage. The relative error decreases with increase in the difference between these two isotope indices used. In a similar study, the investigators (91) have considered uncertainty in the sub-surface components of the lake – not based on the classical propagated error estimation approach – but based on the errors in the conventional (flow-net) method. Therefore, in the present investigation also, a similar approach has been adopted and a conservative estimate of 10% (estimated for the water balance method) is considered as uncertainty in the estimation of sub-surface components by isotope mass balance method.

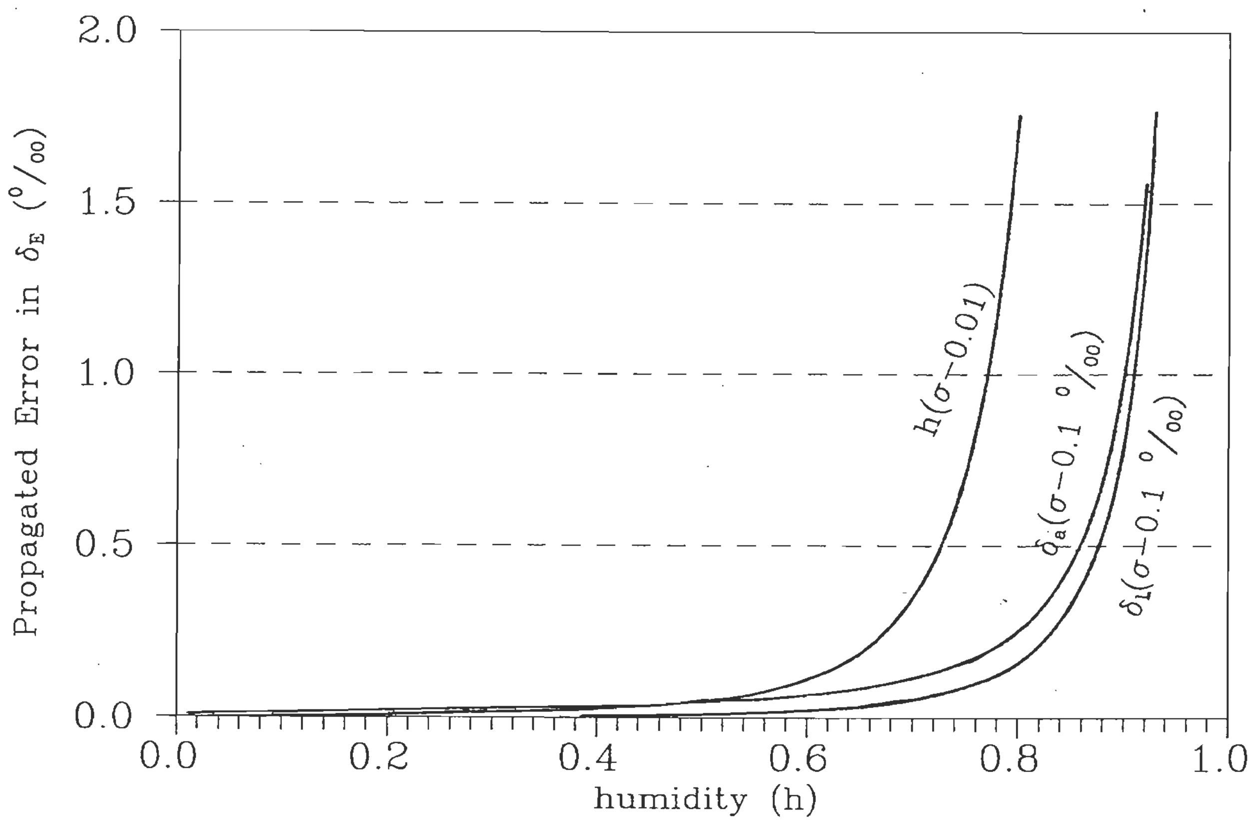


Figure 3.27 Propagated error in δ_E in CGLR model results due to errors in input parameters at different values of h

3.6 Comparison of results

The results presented in Table 3.41 show that the estimates of sub-surface inflow to the lake and outflow from the lake, obtained through isotopic and chemical balance, compare well with those obtained through conventional water balance method.

Table 3.41 SS_I and SS_O data estimated by isotopic, chemical and conventional mass balance methods.

Method of estimation	$\delta^{18}O$		Chloride		Conventional	
	SS_I	SS_O	SS_I	SS_O	SS_I	SS_O
Volume (1000 m^3)	2269	2618	2777	3140	2234	2416
Depth (m)†	5.1	5.88	5.99	6.78	4.82	5.21
% to total inflow or outflow	51	56	55	59	50	54
Lake WRT -Years‡	1.93		1.77		1.92	

† Estimated volumes have been converted into units of depth by normalising to the maximum lake surface area, 463365 m^2.

‡ Lake water retention time has been calculated, assuming mean depth of the lake as 18.52 m. Time difference between the dates of sampling considered was 380 days. Appropriate corrections have been made to calculate the total inflow in 365 days.

Water retention time (WRT) of the lake, is a better parameter to compare the results as WRT is a function of the lake size. WRT computed using isotopic mass balance approach is 1.93 y, chloride mass balance is 1.77 y and conventional water balance is 1.92 y. The results obtained by all the three methods do not vary significantly from each other and compare very well within the error limits. The WRT computed using the isotopic and chloride mass balance approach are more reliable, as they have been derived independently without considering outflow through pumping and springs. The results of isotope and chloride mass balance methods support the conceptual model developed for the Nainital lake.

CHAPTER 4

STUDY OF RIVER GANGA AND GROUNDWATER INTERACTION

4.0 STUDY OF RIVER GANGA AND GROUNDWATER INTERACTION

4.1 Introduction

River Ganga and its canal system, which draw water from the river play decisive role in the agricultural development of the western Uttar Pradesh. Therefore, a better understanding of the nature of interaction between river Ganga and groundwater will help in better management of both groundwater and river. Groundwater interaction with such alluvial rivers may be investigated by analytical and numerical modelling techniques, channel water balance, isotope mass balance and by Dupuit's method using river and groundwater levels.

In this present investigation, river Ganga and groundwater interaction have been investigated along the river course between Hardwar and Narora by both stable isotope and conventional (Dupuit's) methods. Groundwater contribution to the river was estimated using two- and three-component models. Quantification of groundwater discharge to the river at two sites was attempted by using river discharge data. The results of stable isotope method and Dupuit's method are compared and discussed.

The findings of the earlier study, arrived at by water balance method for the period 1967 - 73, indicate that between October and June groundwater contributes to the regeneration of river Ganga from Hardwar to Narora, and it is maximum during October and minimum during April (145). On the other hand, the findings, arrived at by stable isotope technique in the same study area for the period 1982-83, reveals that groundwater component in the river is maximum during April and negligible during August (105).

4.2 Study Area

River Ganga originates from Gangothri glacier located at an altitude of 3142 m above m.s.l. in the Himalaya. After flowing through the hilly terrain for about 260 km, the river enters the plains at Hardwar. The 221 km long river course between Hardwar and Narora has been selected for the present study. The study area falls between 28° and 30° N latitudes & 77°45′ and 78°15′ E longitudes (Figure 4.1). In this reach, river Ganga is joined by smaller rivers like Malin and Chhoya from the east and Solani and Ban Ganga from the west. These small rivers flow only during monsoon season and during rest of the year, they are generally dry.

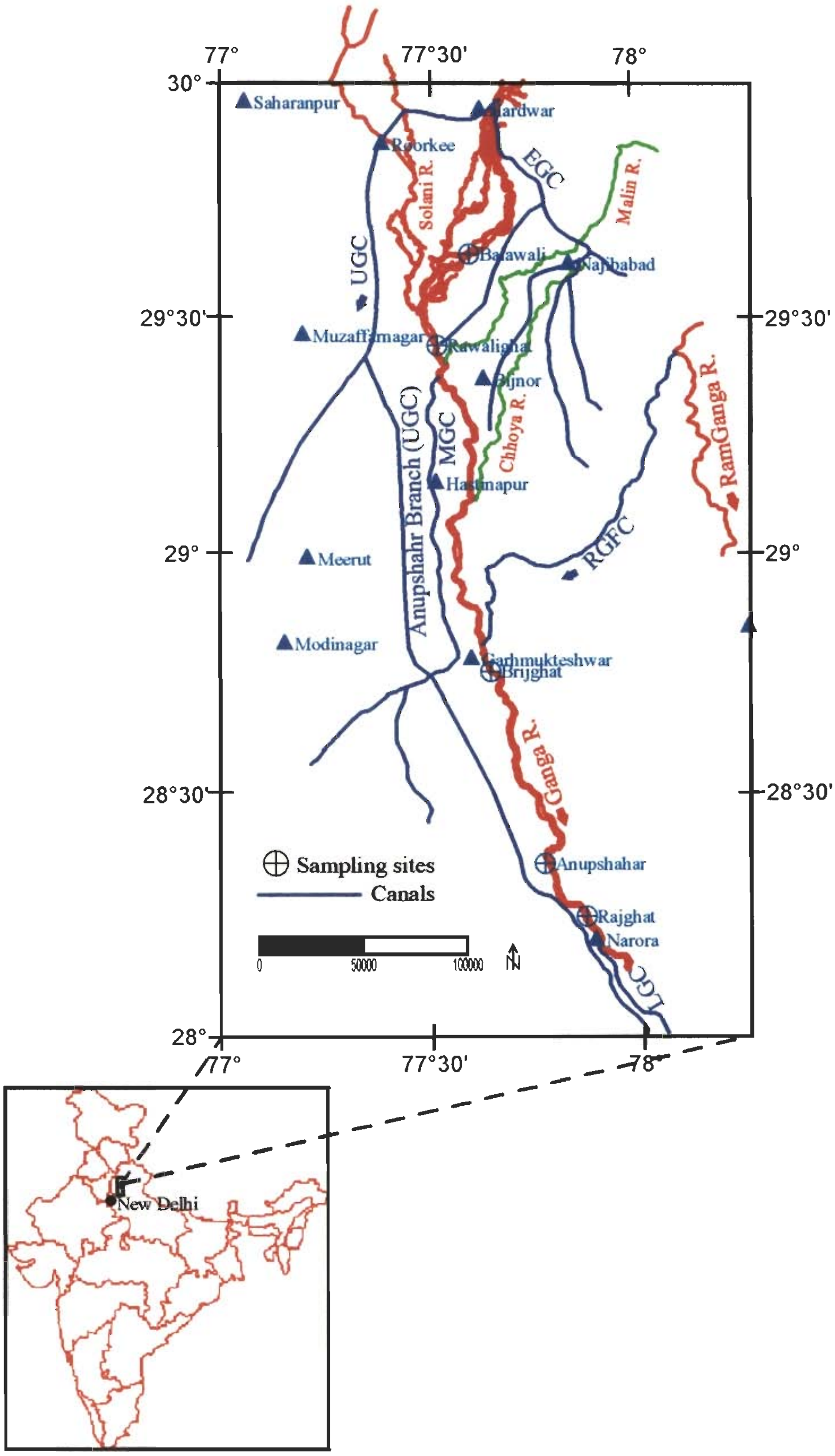


Figure 4.1 Location map of the study area showing river Ganga between Hardwar and Narora.

The selected river reach caters to three major irrigation canal systems viz., the Upper Ganga Canal, Madhya Ganga Canal and Lower Ganga Canal. Recently, Eastern Ganga Canal has also been constructed, which takes off from Hardwar. Notable inflow to river Ganga is Ram Ganga Feeder Canal, which joins river Ganga between Rawalighat and Brijghat. Ram Ganga Feeder Canal flows only during December to May, but is generally dry from the second week of May to November. Since the canal flow takes place in non-monsoon season, which is the lean flow period for river Ganga, the discharge from Ram Ganga Feeder Canal may change the isotope characteristics of the river.

Anupshahr branch of Upper Ganga Canal flows almost parallel to river Ganga on its western side. The distance between Anupshahr branch and the river is about 10 to 13 km. Groundwater level analysis shows that Upper Ganga Canal acts as recharge source for unconfined aquifer between Upper Ganga Canal and the river (144).

The study area is a part of the Indo-Gangetic plains of the Quaternary age. It is made up of unconsolidated fluviatile formation comprising of sand, silt, clay and kankar with occasional bed of gravel. The thickness of alluvium in the Indo-Gangetic plain is roughly 2500 m to 3000 m (144). Geologically, the sediments are favourably embedded for the occurrence of groundwater in major part of the area (156).

Mean annual rainfall in this region is about 990 mm. Most of the rainfall is received during the monsoon season (June to September). Approximately, 25% of the rainfall received, recharges the groundwater (35, 144). Apart from this, the canal systems act as major source of groundwater recharge. The seepage from the canal systems is approximately 1.5 to 2.5 cumecs/10^6 m^2 of wetted perimeter (144). According to the studies conducted in the area, 40% of the canal input reaches the groundwater reservoir as canal seepage and field losses (156).

4.3 Dupuit's Method

Surface water and groundwater interaction may be studied, by employing water level fluctuation data with Dupuit's assumptions. This method is also known as Statistical Method (146). It is applicable to unconfined steady-state alluvial aquifers underlained by impermeable boundary. The method involves analysis of data on groundwater level adjacent to the river and river water level.

4.3.1 Methodology

In order to study the river – groundwater interaction using conventional method, groundwater levels were measured with piezometers installed at Balawali, Rawalighat (near Bijnor), Brijghat (near Garhmukteshwar), Anupshahr and Rajghat (near Narora). The piezometers were installed at distances of 50, 150 and 250 m from the river bank, in a linear pattern and perpendicular to the river course. Monthly groundwater levels were monitored directly and monthly river water levels were computed by using the daily river level data obtained from various government agencies.

4.3.1.1 Analysis of groundwater levels

Observed monthly depth of groundwater table at the piezometer sites, which have been used to calculate the hydraulic gradient (I), which are presented in Table 4.1. The observed water levels are shown in Figures 4.2a to 4.2e. The variation in groundwater levels and the direction of the hydraulic gradient at different sites are discussed below.

Balawali: It is seen from Figure 4.2a, that the groundwater level gradient is towards the river in the non-monsoon months. During June, a reverse trend is observed in the groundwater gradient. However, between August and October the gradient is not well defined, and the groundwater levels do not indicate whether the river was influent or effluent during this period. The water level in the second piezometer is higher than the first and third, thus indicating the presence of a groundwater ridge. It has been observed that groundwater ridges developed parallel and nearer to streams, in response to rainfall in humid areas (136). The groundwater levels observed at Balawali, during monsoon season, are closer to the ground level. Therefore, it is possible that the infiltration of rainfall might have converted the near surface tension saturated capillary fringe into a pressure saturated zone, thus leading to the development of the groundwater ridge.

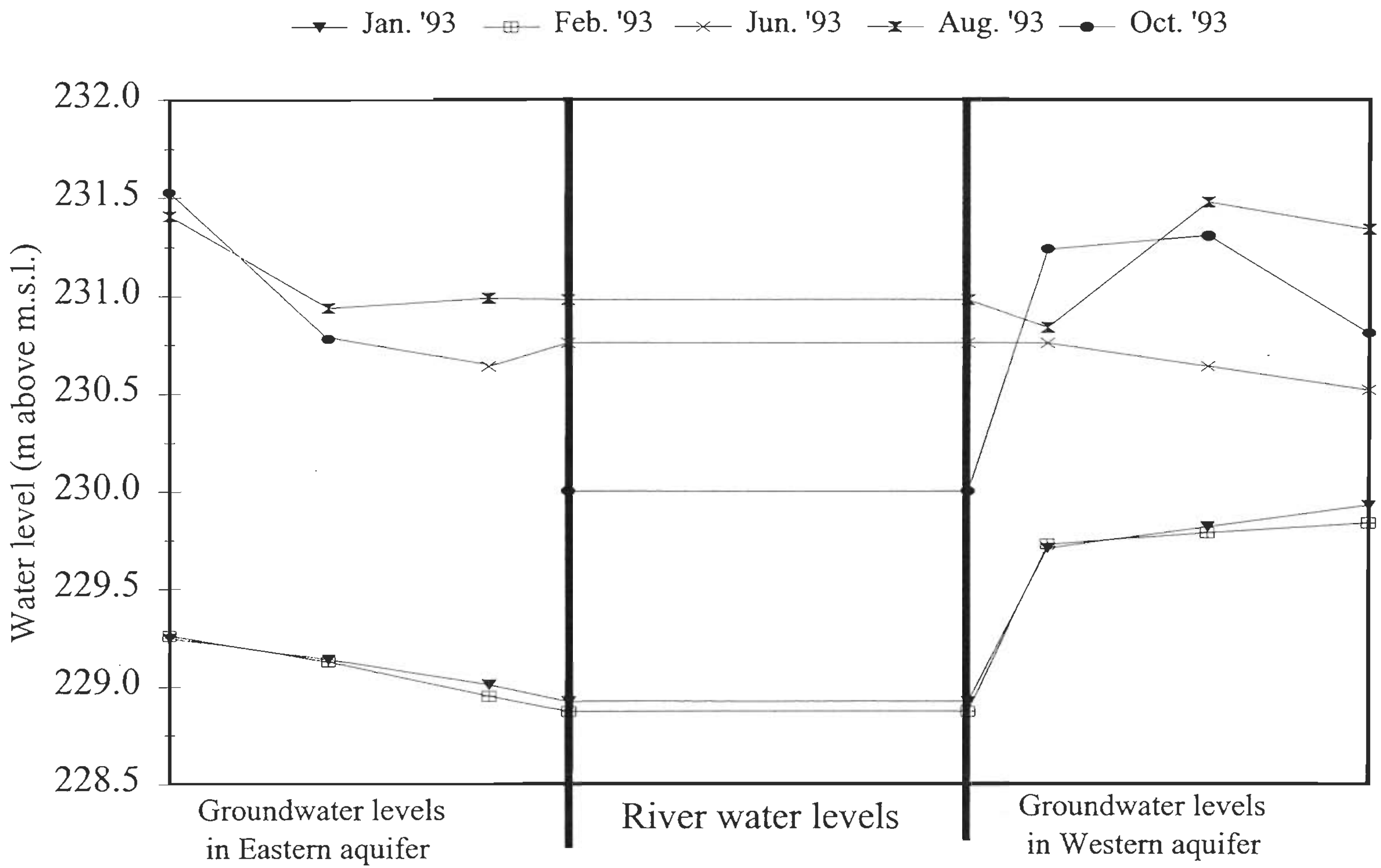


Figure 4.2a. Groundwater levels and Ganga river levels at Balawali

Table 4.1 Hydraulic gradient (I) computed using the water level data observed in the piezometers installed at selected sites along river Ganga during 1993.

Month	Balawali		Brijghat		Anupshahr		Rajghat	
	East	West	East	West	East	West	East	West
Jan.	1.26E-03	1.38E-03	6.0E-04	1.0E-03	9.0E-04	2.1E-03	-9.0E-04	1.5E-03
Feb.	6.32E-04	5.75E-05	6.0E-04	1.3E-03	8.0E-04	2.0E-03	-1.4E-03	2.9E-03
Mar.	9.19E-04	9.73E-04	1.0E-03	8.0E-04	3.0E-04	2.0E-03	-1.5E-03	1.0E-03
Apr.	-	1.95E-03	-	-	-2.4E-03	1.6E-03	-	7.0E-04
May	2.88E-04	4.02E-04	-2.0E-04	-4.5E-03	-2.3E-03	1.8E-03	-	7.0E-04
Jun.	-1.38E-03	-	-5.9E-04	-6.8E-04	-	-	-	-2.0E-04
Jul.	-	-	-1.1E-03	-4.0E-03	-	-	-	-
Aug.	1.66E-03	1.99E-03	1.4E-03	1.8E-03	-	-	-	4.0E-04
Sep.	-	-	-	-	-	-	-	-
Oct.	2.99E-03	5.65E-03	1.9E-03	5.6E-03	-	-	-	1.8E-03
Nov.	2.24E-03	2.50E-03	-	-	-	-	-	-
Dec.	5.14E-03	5.47E-03	-	-	-	-	-	-

* Negative sign indicates influent conditions

Rawalighat: It is seen from Figure 4.2b, that the groundwater level data indicate seepage from the river. During the months of January to March, there are no well defined groundwater level graidents *vis-a-vis* the river levels. However, the river seemed to have gained water from western aquifer and lost into the eastern one. This is due to the presence of Madhya Ganga Barrage, where water level is always maintained. The river water level observed immediately downstream of the Barrage does not represent true conditions and hence it is misleading. The number of piezometers installed are insufficient to evaluate true groundwater gradient in this locality.

Brijghat (Garhmukteshwar): At Brijghat, for the most part of the year, the groundwater level data indicate that the river is gaining water from the aquifer, except during July (Figure 4.2c). Even in the month of July, seepage from the river seems to be retained as bank storage that subsequently flows back to the river as seen from the data of August 1993.

Anupshahr: The groundwater levels observed in the western aquifer closer to the river indicate that the groundwater gradient is always towards the river, except during the month of July

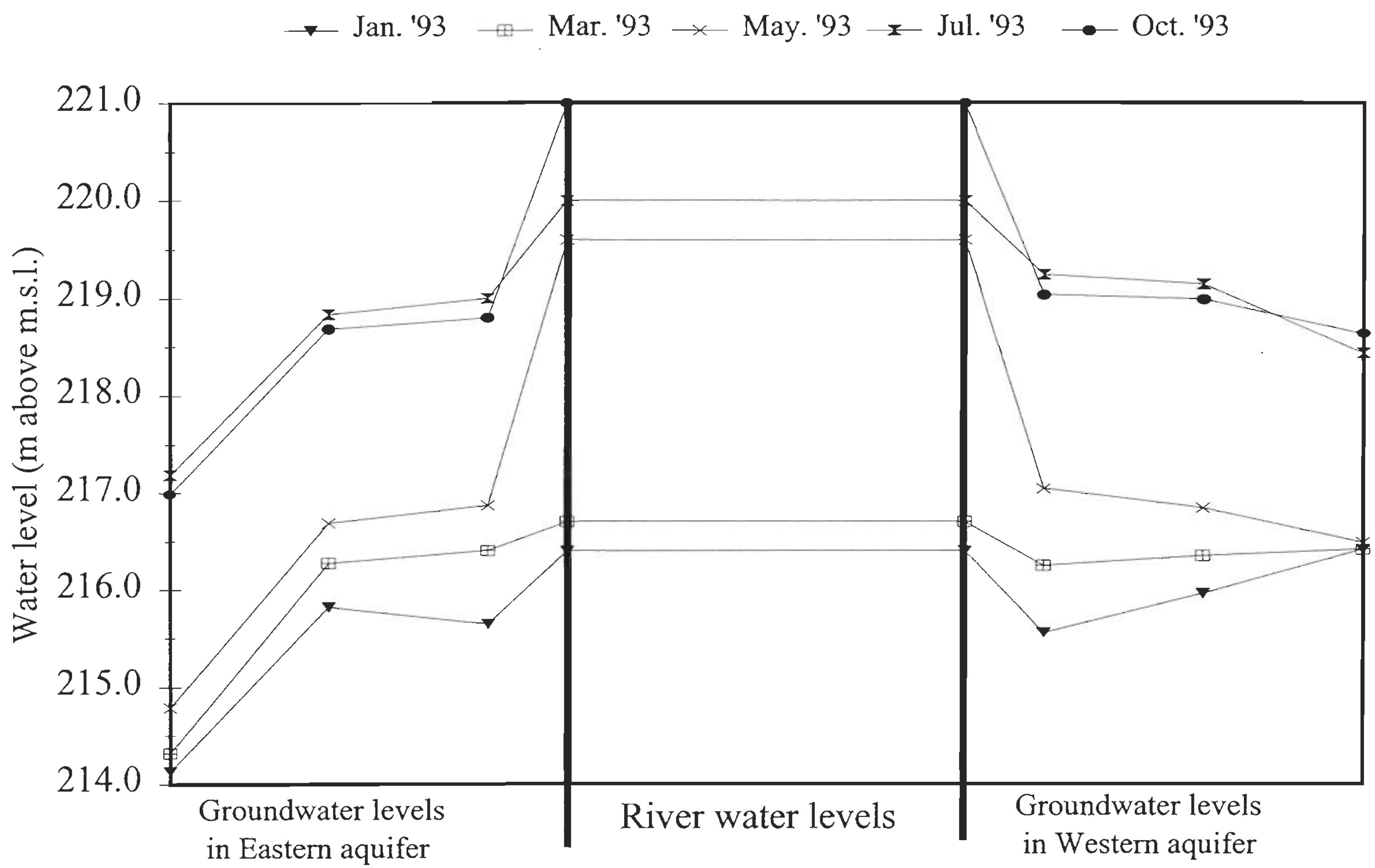


Figure 4.2b. Groundwater levels and Ganga river levels at Rawalighat

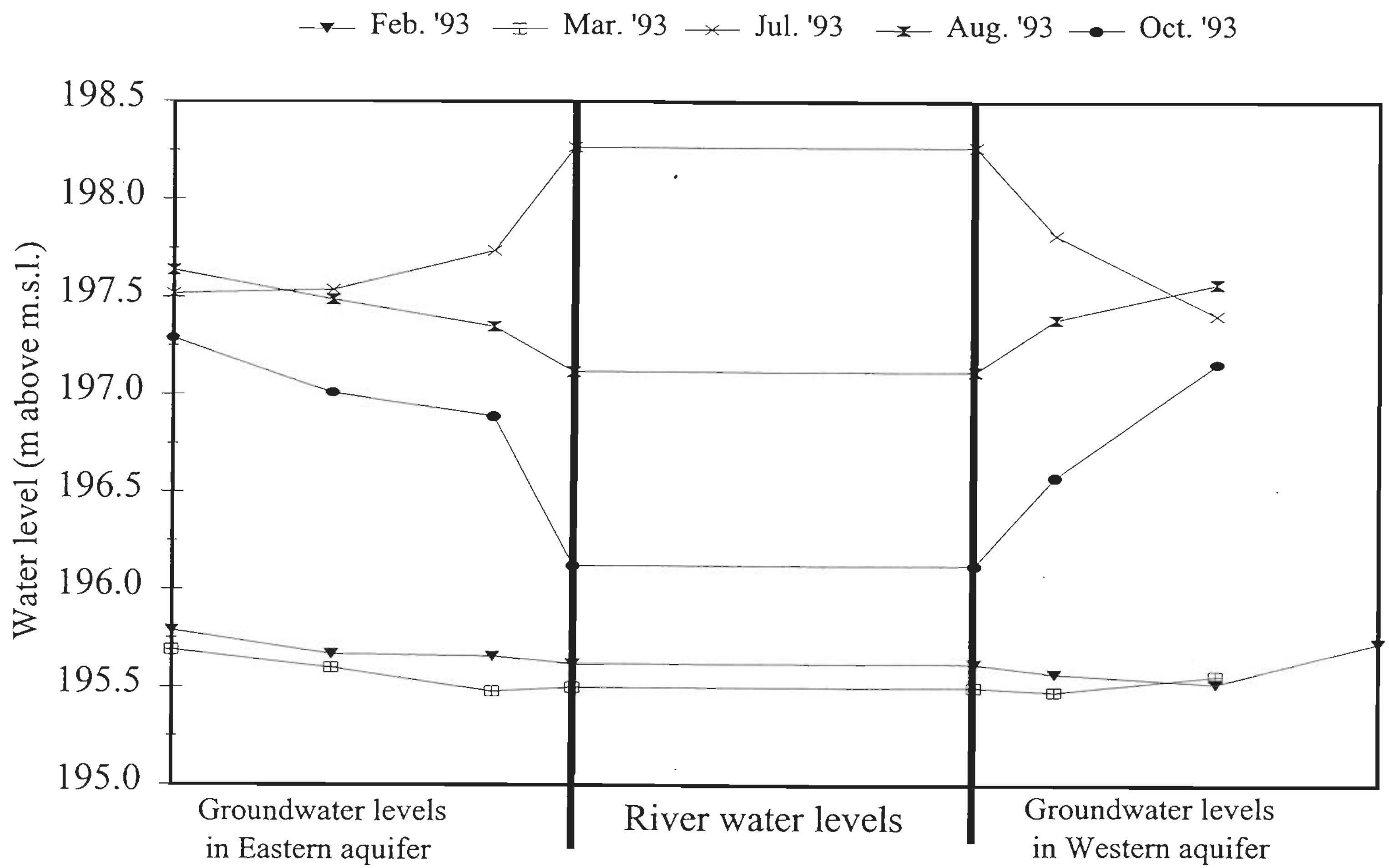


Figure 4.2c. Groundwater levels and Ganga river levels at Brijghat

(Figure 4.2d). The effect of river seepage is confined to a maximum distance of approximately 150 m from the river bank in the western side.

Rajghat (Narora): At Rajghat (Narora) the groundwater gradient is towards the river in the western aquifer and away from it in the eastern aquifer during non-monsoon season (Figure 4.2e). Therefore, at Rajghat the river does not determine the base level of groundwater.

In a similar study, on the river course between Hardwar and Rawalighat (146), it has been noted that river Ganga is effluent on one bank and influent on the other bank, along the course between Ajitpur and Tatwala (both sites are located between Hardwar and Balawali). The following inferences are drawn based on the above observations:

a) river Ganga does not significantly recharge the aquifers on either side of the river during monsoons, and the seepage from the river is retained as bank storage, which subsequently flows back into the river,

b) the river does not determine the base level of the regional groundwater table at least in selected zones, and

c) at Rawalighat the Madhya Ganga Canal reservoir recharges the groundwater throughout the year. The recharged water may flow towards the river in the downstream direction. However, more well points are needed to evaluate the extent of groundwater recharge / discharge.

4.3.1.2 Estimation of hydraulic conductivity

The hydraulic conductivity (K) at the piezometer sites have been estimated by two methods, viz. the particle size analysis and tracer dilution methods. Sediments were collected during the installation of the piezometers, for particle size analysis. Samples at every 1-meter depth were sieved and particle size distribution was evaluated. The results show that the sediments are generally of medium to fine sand with little silt/clay. Adopting Hazen's method (49), K - values were calculated and the results are given in Table 4.2. Hydraulic conductivity values of unconfined aquifer from over 200 sites in upper Ganga basin (Western Uttar Pradesh) based on column drainage and particle size distribution illustrates that the K- values show large spatial variations in the study area (104).

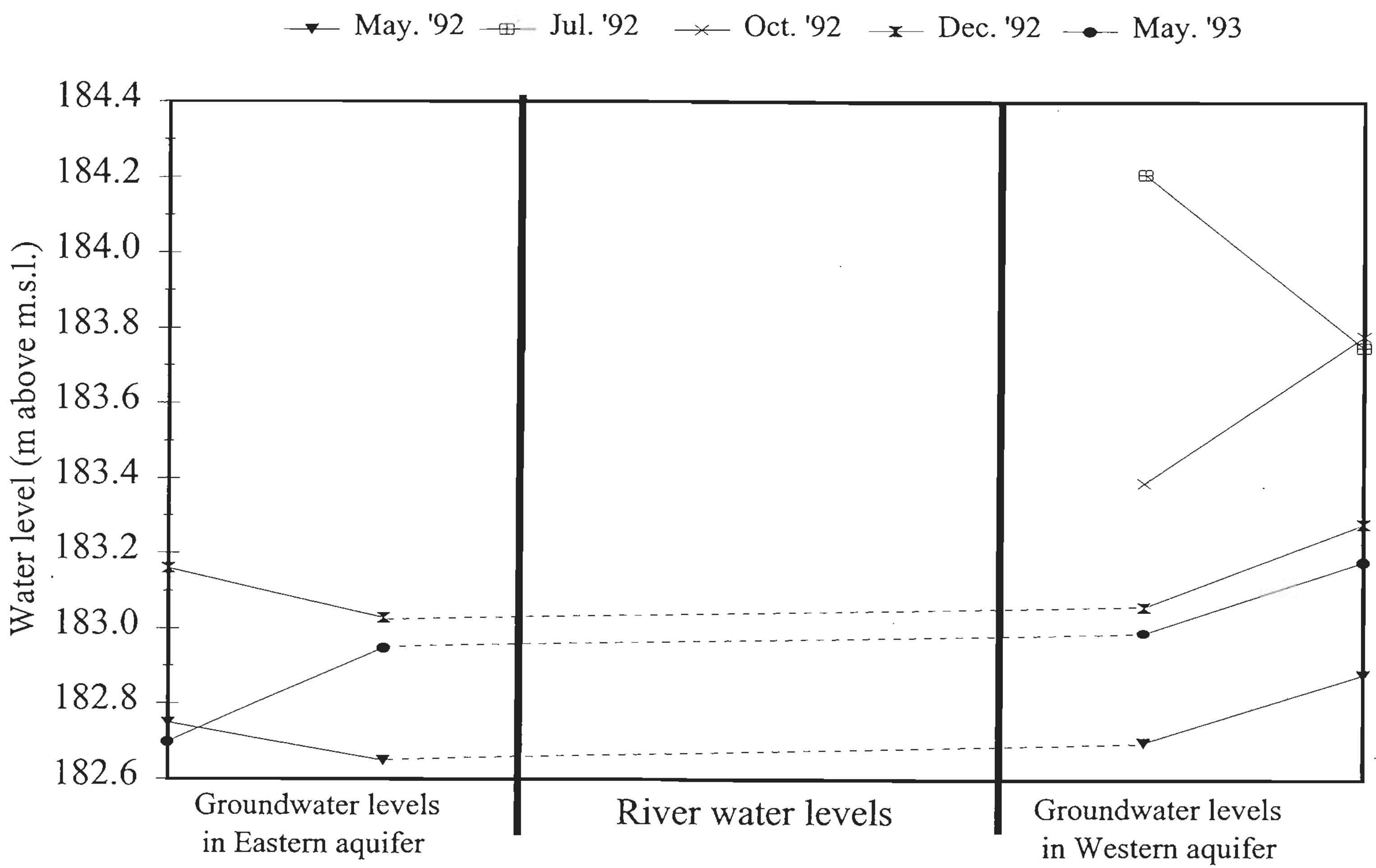


Figure 4.2d. Groundwater levels and Ganga river levels at Anupshahr
[River water levels at Anupshahr were not available]

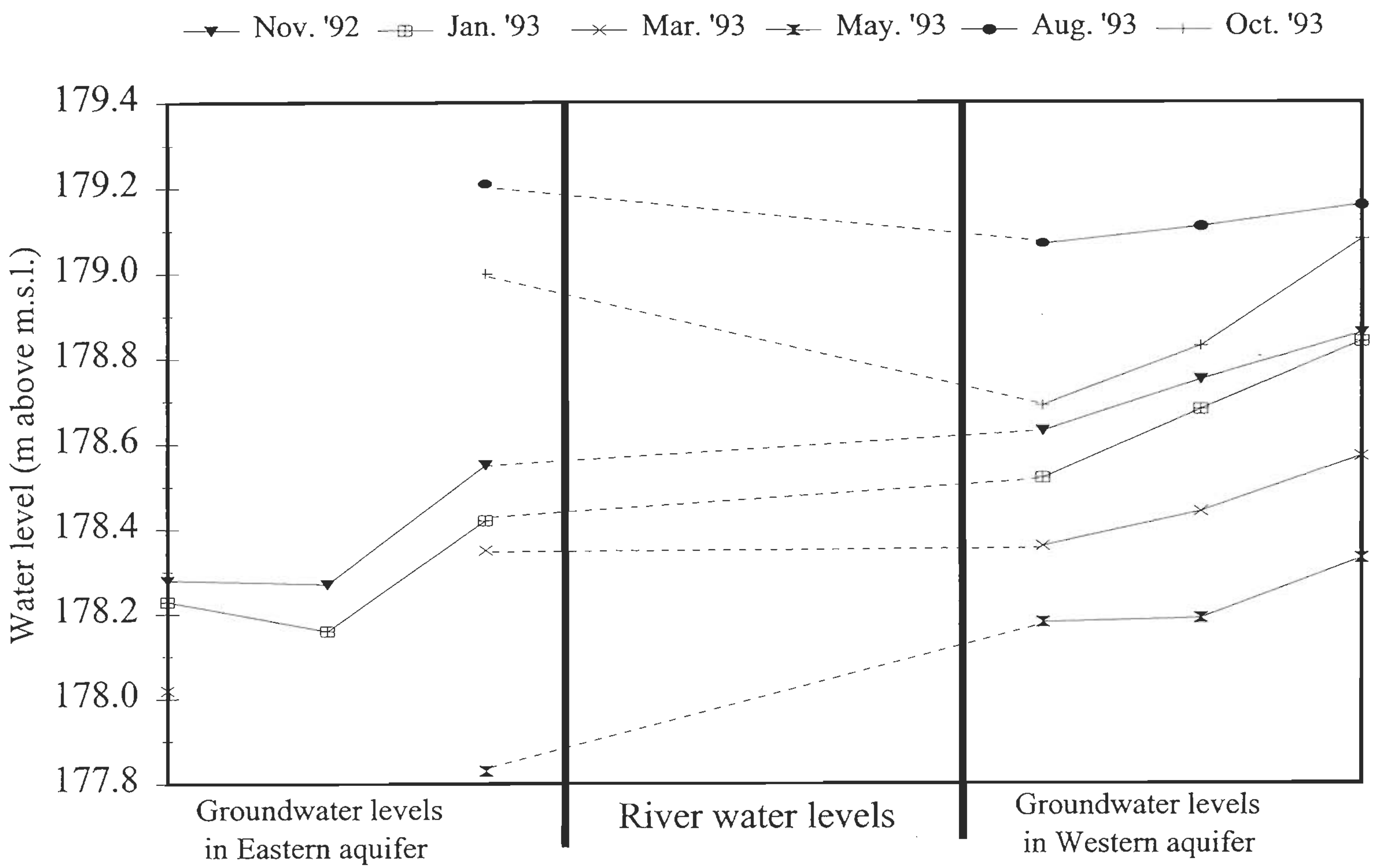


Figure 4.2e. Groundwater levels and Ganga river levels at Rajghat
[River water levels at Rajghat were not available]

Hazen method utilises a coefficient that has a large range. Therefore, uncertainty in the results is also relatively larger. Hence, the single well tracer dilution technique was attempted for the estimation of aquifer hydraulic conductivity in the study area. The use of the single well tracer dilution technique has a definite advantage over the conventional methods, and it is well-established (45, 13). It involves introduction of known quantity of tracer in the well and monitoring of the tracer concentration over a time period. The filtration velocity (V_f) in m/d can be estimated using the following relationship (45)

$$V_f = \frac{\pi r_1}{2\alpha t} \ln\left(\frac{C_0}{C}\right) \tag{4.1}$$

where, r_1 is the internal radius of the well (m), t is time (days), C_0 and C are the tracer concentrations at time t = 0, and at time = t. α is the correction factor to eliminate the distortion of groundwater flow lines caused by the presence of the well. For the wells constructed with plastic screens and without gravel pack, α can be calculated by the following equation (46).

$$\alpha = \frac{4K_1}{K_1\left[1 + \left(\frac{r_1}{r_2}\right)^2\right] + K\left[1 - \left(\frac{r_1}{r_2}\right)^2\right]} \tag{4.2}$$

where, K_1 is the hydraulic conductivity of the well screen, r_1 and r_2 are the internal and external radii of the well screen respectively. The uncommon term K_1 (cm s^{-1}) can be estimated as 0.1f, where, f is the percent perforation of the well screen(46). It can be seen that calculation of α requires prior information of K, the hydraulic conductivity of the aquifer. However, the influence of K on α will be negligible when $K << K_1$. The method can be used to estimate K - value of the aquifer by using Darcy's equation ($V_f = KI$), if the hydraulic gradient (I) near the well is known.

Tracer dilution experiments were conducted at Balawali, Rawalighat, Brijghat and Rajghat for determining aquifer hydraulic conductivity. Known quantities of ^3H and NaCl solutions were introduced into the wells to label the whole column of water. A special device was fabricated and used to mix the water column thoroughly, and also to periodically collect the water samples from the well. The samples were analysed for tritium using a liquid scintillation counter, and for NaCl concentration by electrical conductivity method.

The values of ln (C_0/C) calculated from the tracer concentration versus time plots are given in Table 4.2. All the wells used in the experiments had the same construction geometry, and therefore a linear relationship between α and K was obtained to facilitate quicker estimation of α - values:

$$\alpha = -0.00039K + 2.166 \tag{4.3}$$

It is clear from equation (4.3) that α - values are not sensitive to K - values, when K_l >> K. α - values range from 2.16 to 2.17 for K - values from 0.0001 to 25 m d^{-1} in the present study. Estimation of hydraulic conductivity (K_p) values at different sites was made using the particle size data by Hazen's method (49). The estimated V_f were then used in the Darcy's equation (V_f = KI) to back calculate K - values. By iterative procedure, correct values of K were determined. The K - values, estimated by both tritium and chemical tracer methods, are shown in Table 4.2.

It is seen from Table 4.2 that Hazen's method underestimates the K - values in the present case. It is also seen that the K - values determined by NaCl tracer (K_N) are lower than those determined by ^3H tracer (K_T). Since tritium is a part of water molecule and a perfect tracer for groundwater movement, the results of tritium dilution can be considered as a better method to determine K.

4.3.2 Estimation of Specific Discharge

It is well known that the specific discharge in an unconfined aquifer can be estimated using Dupuit's equation (49):

$$q' = \tfrac{1}{2} K [(h_1^2 - h_2^2)/L] \tag{4.4}$$

where, q' is specific discharge per unit width of the aquifer (m^2/d), K is hydraulic conductivity (m/day), h_1 is head at the origin (m), h_1 is the head at a distance L along the flow direction (m) and L is the flow length (m). The above equation may be assumed as applicable to the study area and as valid under the following conditions:

a. The hydraulic gradient is equal to the slope of the water table (Dupuit's assumption)

b. The aquifer thickness is 100 m and is underlained by an impermeable layer (consistent with the assumptions made by earlier investigators. (103, 144)

c. The flow in the aquifer is one-dimensional and is perpendicular to the river

Having the above assumptions, groundwater head it the piezometers was computed from the depth to groundwater level data. Specific discharge at different sites was computed by using the above equation and the results are presented in Table 4.3.

4.4 Stable Isotope Method

The utility of stable isotope technique for studying the river and groundwater interaction problems is fairly well established. However, as mentioned earlier in Chapter 2.0, very few studies were conducted on large alluvial rivers. Earlier attempt using isotope technique, for this region was mainly a semi-quantitative analysis to understand the contribution of groundwater to the river (105). The investigators considered the most enriched isotope value at Garhmukteshwar site as the groundwater index for the entire study area. This was supported by higher groundwater levels than the river levels throughout the year at Garhmukteshwar during the study period 1982-83. However, the study did not consider the bank storage mechanism and the temporal and spatial variation in the groundwater isotopic composition. In view of this, both the temporal and spatial isotopic variations of groundwater in relation to river have been focused upon in this present investigation.

Stable isotope values of water samples collected from river Ganga and from the eastern and western aquifers during the period May 1992 to May 1995, during January 1994 and during March, 1993 have been presented in Tables 4.4a, 4.4b and 4.4c respectively.

Controlling Factors: River Ganga as well as aquifers on either side of the river (Table 4.4a), show considerable variation in their stable isotopic composition. The stable isotope values of the river may be affected by the depletion in isotopes due to snowmelt/precipitation at higher altitudes and by the enrichment due to mixing of tributary inflow/groundwater contribution in

Table 4.2 Hydraulic conductivity values determined by Hazen's method and tracer dilution techniques for different sites along river Ganga.

Site	Well #	Kp (m/s)	Tritium tracer			NaCl tracer		
			[ln(C0/C)]t	V_f (m/s)	K_T (m/s)	[ln(C_0/C)]/t	V_f (m/s)	K_N (m/s)
Balawali	BR1	1.93E-05	3.40E-05	9.25E-07	1.93E-04	-	-	-
Rawalighat	RL1	2.36E-05	5.52E-05	1.50E-06	2.06E-04	1.06E-05	1.20E-07	1.60E-05
	RR1	2.36E-05	5.05E-05	1.37E-06	1.58E-04	3.21E-05	4.12E-07	4.69E-05
Brijghat	BL1	2.16E-05	3.09E-05	8.42E-07	2.10E-04	7.72E-06	1.51E-07	3.63E-05
	BR1	2.16E-05	3.77E-05	1.02E-06	1.46E-04	3.21E-05	9.72E-07	1.39E-04
Rajghat	NL1	2.11E-05	7.46E-05	2.03E-06	3.38E-04	6.81E-05	1.31E-06	2.13E-04
	NR1	2.11E-05	4.77E-05	1.30E-06	2.60E-04	3.48E-05	6.45E-07	1.31E-04

Table 4.3 Calculated specific discharge (m^2/day) at different sites for 1993.

Months	Balawali		Brijghat		Anupshahr		Rajghat	
	East	West	East	West	East	West	East	West
January	2.40	2.61	1.13	1.27	1.69	2.70	-2.59	3.36
February	1.20	0.11	1.13	1.63	1.52	2.46	-3.95	6.60
March	1.75	1.85	1.83	0.96	0.51	2.58	-4.49	2.20
April	-	3.70	-	-	-4.39	1.99	-	1.57
May	0.55	0.76	-0.43	-5.67	-4.22	2.23	-	1.57
June	-2.63	-	-1.05	-0.85	-	-	-	-0.52
July	-	-	-1.95	-5.04	-	-	-	-
August	3.16	3.78	2.57	2.21	-	-	-	0.95
September	-	-	-	-	-	-	-	-
October	7.10	13.42	3.53	7.08	-	-	-	4.10
November	5.33	5.94	-	-	-	-	-	-
December	12.20	12.99	-	-	-	-	-	-

Table 4.4a δ[18]O (‰) data of water samples collected during May 1992 to May 1995 from river Ganga, western aquifer (WA) and eastern aquifer (EA) and Upper Ganga Canal (UGC).

Period	Location	River	WA	EA
05/92	Hardwar	-9.1	-8.9	-
	Rawalighat	-8.2	-7.1	-9.3
	Brijghat	-7.2	-5.3	-9.0
	Anupshahr	-7.1	-7.2	-
	Rajghat	-7.3	-7.3	-7.7
11/92	Hardwar	-10.6	-	-
	Balawali	-10.3	-9.2	-8.6
	Rawalighat	-9.7	-	-10.3
	Brijghat	-9.1	-5.7	-9.3
	Anupshahr	-9.0	-8.8	-
	Rajghat	-8.2	-7.5	-7.5
12/92	Balawali	-11.0	-9.5	-7.8
	Rawalighat	-	-10.3	-11.0
	Brijghat	-	-5.0	-9.7
	Anupshahr	-9.4	-8.2	-7.7
	Rajghat	-7.9	-8.3	-7.1
01/93	Balawali	-9.4	-8.9	-8.0
	Rawalighat	-9.3	-10.7	-9.5
	Brijghat	-7.2	-4.7	-8.9
	Anupshahr	-7.0	-7.2	-8.6
	Rajghat	-7.8	-	-7.6
02/93	Hardwar	-10.3	-10.0	-
	Balawali	-9.6	-8.7	-7.9
	Rawalighat	-9.5	-10.0	-9.8
	Brijghat	-9.0	-4.7	-
	Anupshahr	-8.4	-8.0	-8.0
	Rajghat	-7.4	-7.6	-7.5
03/93	Balawali	-6.6	-8.2	-4.3
	Rawalighat	-4.5	-7.7	-10.8
	Brijghat	-6.7	-5.1	-5.6
	Rajghat	-7.0	-7.4	-6.4
06/93	Hardwar	-8.8	-6.5	-4.6
	Balawali	-8.0	-4.1	-6.3
	Rawalighat	-8.6	-	-9.0
	Brijghat	-7.8	-4.2	-3.9
	Anupshahr	-8.7	-	-
	Rajghat	-8.1	-	-

Period	Location	River	WA		EA	
08/93	Hardwar	-10.4	-9.7		-	
	Balawali	-10.8	-6.7		-6.1	
	Rawalighat	-10.1	-10.1		-12.2	
	Brijghat	-9.4	-8.1		-8.8	
	Anupshahr	-8.9	-7.4		-	
	Rajghat	-8.1	-8.4		-9.4	
01/94	Hardwar	-8.4	-8.8		-	
	Balawali	-4.6	-		-6.4	
	Rawalighat	-7.7	-7.7		-9.7	
	Brijghat	-6.3	-4.7		-	
	Anupshahr	-6.9	-9.1		-	
	Rajghat	-8.3	-7.0		-6.4	
03/94	Hardwar	-10.0	-		-	
	Balawali	-6.7	-8.2	-7.2	-	-4.7
	Rawalighat	-7.8	-6.6		-8.4	
	Brijghat	-6.2	-4.4	-8.5	-8.8	-7.4
	Anupshahr	-8.4	-6.8	-8.4	-	
	Rajghat	-7.6	-5.7		-9.4	
08/94	Balawali	-12.4	-		-5.4	
	Rawalighat	-	-7.0		-	
	Brijghat	-	-3.7		-5.7	
	Anupshahr	-9.6	-8.3	-9.8	-	
	Rajghat	-8.0	-5.2		-6.8	
10/94	Hardwar	-	-7.1		-6.4	
	Balawali	-6.4	-7.7		-	
	Rawalighat	-5.7	-	-8.2	-	-5.5
	Brijghat	-	-5.0		-	-7.0
	Anupshahr	-	-	-7.2	-	
	Rajghat	-	-	-8.4	-	-6.9
	Gangesri				-5.2	
05/95	Hardwar	-11.6	-10.6		-	
	Balawali	-10.8	-10.2		-10.4	
	Rawalighat	-10.9	-9.8		-11.4	
	Brijghat	-8.1	-7.2		-8.2	
	Anupshahr	-11.0	-7.2		-	
	Rajghat	-11.3	-8.9		-9.6	
	UGC	-9.7				

Table 4.4b δ^{18}O (‰) data of samples collected during January 1994 with distances of sampling from the river Ganga. Profiles 5 and 26 corresponds to Balawali and Rajghat (Narora) respectively.

Profile #	Western Aquifer			River	Eastern Aquifer		
	10.0 km	5.0 km	1.0 km		1.0 km	5.0 km	10.0 km
2		Dhanpura -7.9	Shahpur -6.5				
3	Sultanpur -6.7	Bakarpur -6.9	Jaspur Ranjitpur -6.8		Ganga Mandir - 4.5	Saiyadapuri -6.8	Mondawali -7.4
5	Laksar -6.9			-4.6	Balawali RS -7.0	Chandok Gaon -8.8	Pundari Kalan -6.0
7				-4.6			
10	Katiya -8.0	-	-	-7.7	-	-	Jandarpur -7.5
13				-7.2			
14		Mamipur -6.6	Mukhdampur -7.2			Theth -8.4	Dhakiya (Mohibilapur) -5.2
18		Guttiya (LKO crossing) -8.4	Main rd check post -3.1	-6.3			Gajraula Junction -7.7
20				-6.5			
21	Amarpur -8.1	Yunaspur -9.2	Mawai -9.4	•	Biharipur -10.6	Pathra -7.9	Gangesri -6.6
23	Barauli -8.6	Karanpur Nagla -8.4	Anupshahr (outer) -6.9	-6.4		Tomi -7.8	Bahat Karan -7.8
26				-8.3			

the plains and also by evaporation. Similarly, in the groundwater the isotopic characteristics may change due to a) rainfall recharge, b) canal recharge, c) river recharge, d) Irrigation return flow and e) evaporation from the groundwater. The first factor determines the overall isotopic composition of the groundwater, the next two factors may show depletion in the isotopic values while the last two factors show enrichment.

The isotope data (Table 4.4a) reflect the influence of these factors in varying degree in the study area. It is observed that at some sites, the influence of irrigation return flow is more prominent during certain season, whereas that of the canal seepage during other seasons. In the following sections, variations in the isotopic characteristics of river Ganga, groundwater on either side of the river, and rainfall in the region are discussed.

4.4.1 Stable Isotope Composition of Rainfall

From the IAEA/WMO GNIP data, the $\delta^{18}O$ - δD relationship of rainfall at New Delhi station (located very close to the study area; Figure 4.1) is found to be:

$$\delta D = (7.9 \pm 0.3) \cdot \delta^{18}O + (7.6 \pm 2.0) \quad [r = 0.96;\ n = 27] \tag{4.4}$$

Equation 4.4 was derived using the weighted average values of the monsoon season only. Since the rainfall during non-monsoon season is negligible, only the monsoon rainfall is likely to influence the isotopic characteristics of groundwater. The Meteoric Water Line derived for New Delhi is very close to Global Meteoric Water Line (125). The rainfall isotopic data of New Delhi station show that the long-term monsoon weighted averages of $\delta^{18}O$ and δD are -6.4‰ and -43‰ respectively. Therefore, it is likely that the $\delta^{18}O$ of the groundwater in the study area may be about -6.4‰.

Table 4.4c Stable isotopic composition of samples collected at Rajghat (Narora) during March, 1993

Location	Sample No.	$\delta^{18}O$ (‰)	δD (‰)
Western Aquifer	RB1	-7.4	-52
	RB2	-	-52
	RB3	-7.3	-51
Eastern Aquifer	LB1	-8.3	-57
	LB2	-6.4	-51
River Ganga	Western extreme	-6.9	-49
	Eastern extreme	-7.0	-49

4.4.2 Oxygen isotope composition of groundwater

Frequency distribution analysis (with a class interval of 1.0‰) of the stable isotope data of groundwater samples collected closer to the river, during pre-monsoon season (Figure 4.3a), shows that the peaks of the western (-8.5‰) and eastern (-9.5‰) aquifers do not coincide. This indicates that the processes that influence the groundwater isotopic characteristics, in the eastern and western aquifers, are different. This is also confirmed by monsoon and post-monsoon isotopic data (Figure 4.3b). However, as compared to the pre-monsoon data, the monsoon data show $\delta^{18}O$ enrichment.

Analysis of groundwater samples, collected at larger distances from the river during January 1994 (Figure 4.3c), indicate that the frequency distribution of $\delta^{18}O$ in the eastern aquifer is Gaussian and the peak is about -7.5‰. On the other hand, the frequency distribution of $\delta^{18}O$ in the western aquifer is similar to Figure 4.3a, substantiating that the spatial variation in the $\delta^{18}O$ of the aquifers is considerable and is brought about by a variety of processes. The western aquifer has probably been influenced by the recharge from Upper Ganga Canal and its Anupshahr branch and from irrigation return flow. On the other hand, the eastern aquifer seems to have been influenced by Eastern Ganga Canal in the northern part of the study area, Ram Ganga Feeder Canal, irrigation return flow and in certain zones by river Ganga. It is also noticeable that due to better canal network and irrigation facilities available in the western side of the river, the contribution of irrigation return flow may be higher to the western aquifer as compared to the

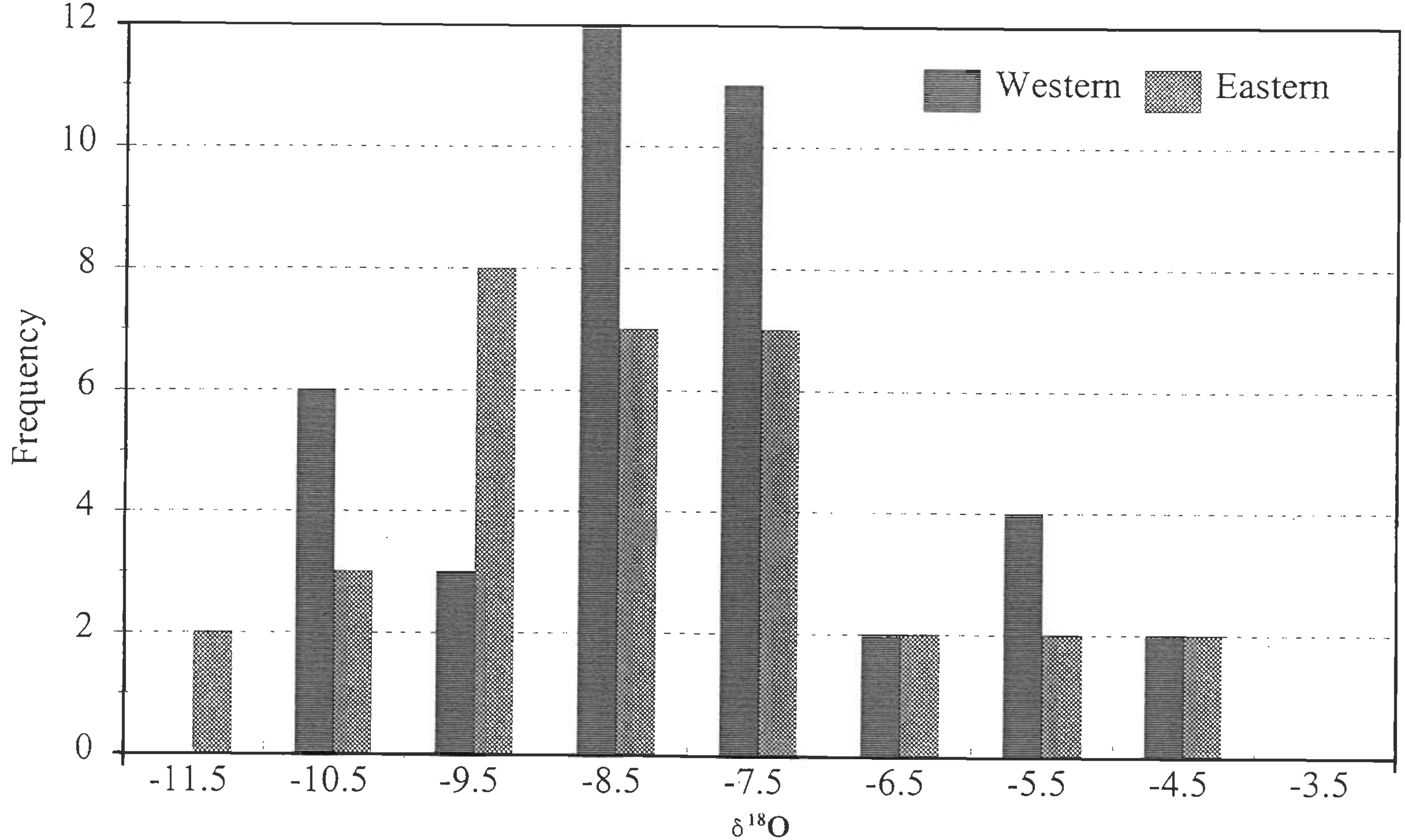


Figure 4.3a Histogram showing the $\delta^{18}O$ values of groundwater samples collected closer to the river during pre-monsoon season.

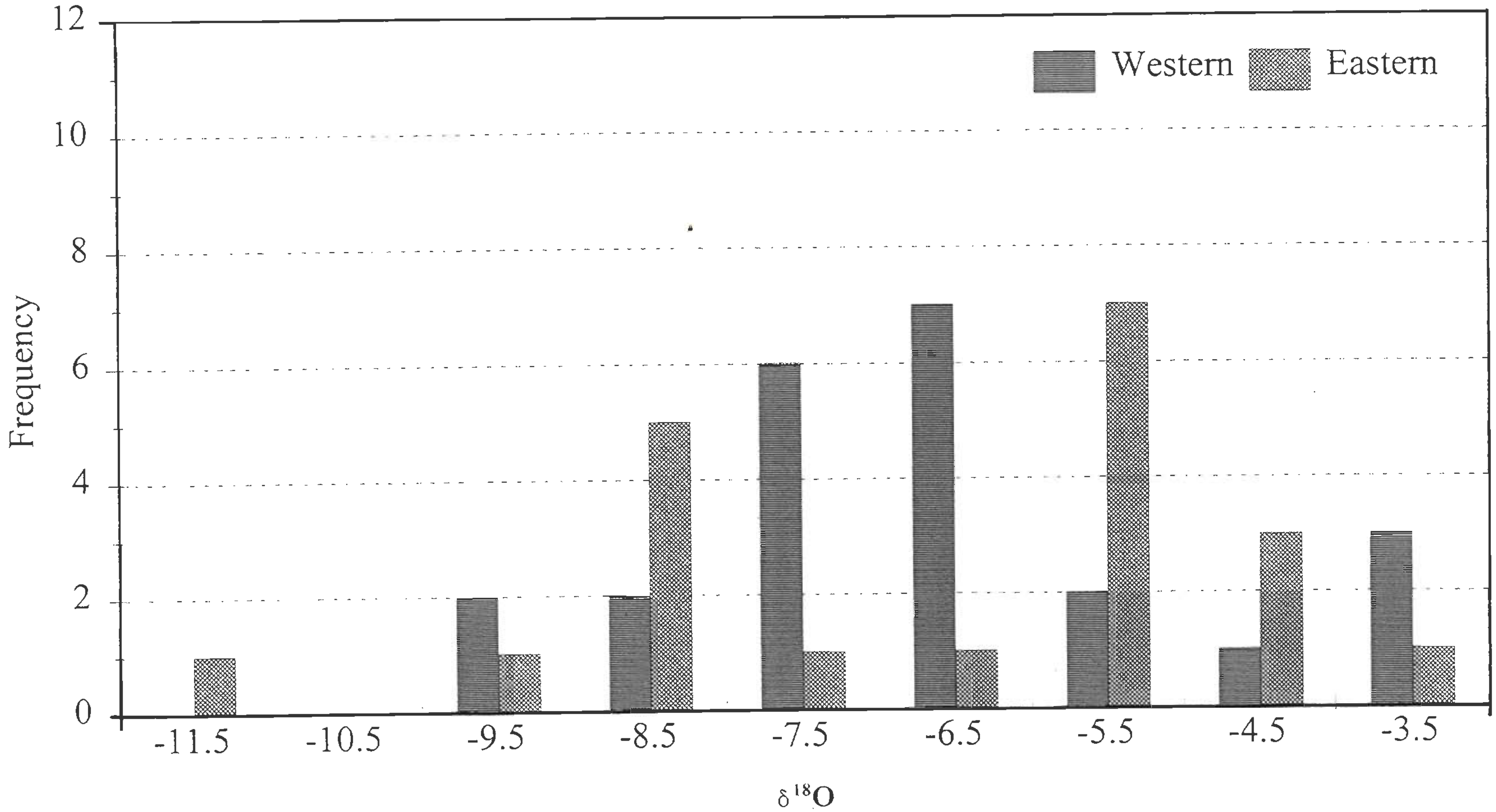


Figure 4.3b Histogram showing the $\delta^{18}O$ values of groundwater samples collected closer to the river during monsoon and post-monsoon seasons.

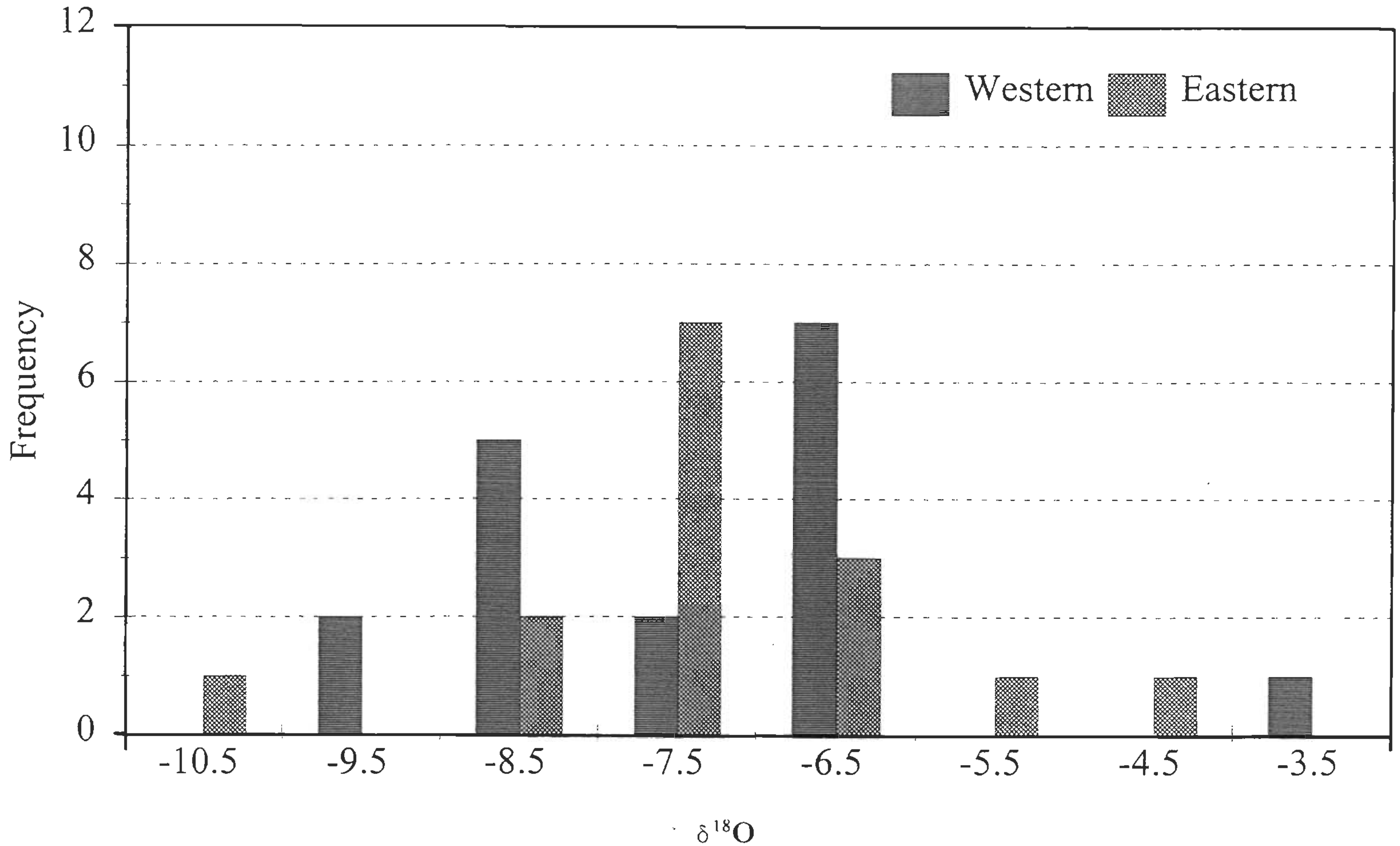


Figure 4.3c Histogram showing the $\delta^{18}O$ values of groundwater samples collected 1.0 to 10.0 km away from river Ganga during January, 1994.

eastern aquifer. The mean isotopic values at different seasons for western and eastern aquifers have been computed and presented in Table 4.5.

Table 4.5 Mean $\delta^{18}O$ of samples collected from May, 1992 to May, 1995 at different sites.

Sites		Pre-Monsoon	Monsoon	Post-Monsoon
Eastern Aquifer	Hardwar	-	-4.6	-
	Balawali	-8.2	-5.9	-9.1
	Rawalighat	-9.7	-10.6	-10.3
	Brijghat	-9.2	-6.1	-7.2
	Anupshahr	-8.3	-	-8.2
	Rajghat	-8.1	-8.1	-7.9
Western Aquifer	Hardwar	-9.6	-8.1	-
	Balawali	-9.0	-5.4	-8.5
	Rawalighat	-8.7	-8.6	-11.0
	Brijghat	-6.5	-6.2	-7.7
	Anupshahr	-7.6	-7.4	-8.3
	Rajghat	-7.3	-6.8	-7.3
River Ganga	Hardwar	-9.9	-9.6	-10.6
	Balawali	-9.1	-9.4	-10.7
	Rawalighat	-8.9	-9.4	-9.7
	Brijghat	-7.3	-8.5	-9.1
	Anupshahr	-8.0	-9.1	-9.2
	Rajghat	-8.1	-8.2	-8.1

Balawali: $\delta^{18}O$ values of the western aquifer at Balawali are comparable during pre- and post-monsoon seasons (-9.0‰ & -8.5‰) and relatively depleted as compared to the monsoon season (-5.4‰). Similar to the western aquifer, the eastern aquifer also shows a relatively depleted $\delta^{18}O$ values during pre- and post-monsoon seasons (-8.2‰ and -9.1‰) as compared to the monsoon season (-5.9‰). Since river Ganga (-9.4‰) is much more depleted than the groundwater during monsoon season, the enrichment in $\delta^{18}O$ may be due to rainfall recharge and also from the evaporation of groundwater. The evaporation of the groundwater during monsoon season is possible, as the water level is much closer to the ground surface (less than one metre).

Between Hardwar and Balawali, the western aquifer had $\delta^{18}O$ content that ranged from -6.5‰ to -7.9‰ (sampling profiles 2 and 3; Table 4.4b) with an average of -7.0‰, which is slightly lesser than the long-term average rainfall index (-6.4‰). $\delta^{18}O$ in the eastern aquifer ranges from -4.5‰ to -8.8‰ (Table 4.4b), with an average of -6.9‰. Nevertheless, $\delta^{18}O$ becomes more negative, when moving away from the river towards east. This may be due to recharge that occurs from the Eastern Ganga Canal, which offtakes from Hardwar. It clearly indicates that the groundwater in Balawali area does not have unique $\delta^{18}O$ value, but varies both temporally and spatially.

Rawalighat: In this locality $\delta^{18}O$ values (Table 4.5) of pre- and post-monsoon seasons both western (-8.7‰ and -11‰) and eastern (-9.7‰ and -10.3‰) aquifers are comparable to monsoon season (western aquifer -8.6‰ & eastern aquifer -10.6‰). It indicates that the evaporation process has not influenced the groundwater isotopic characteristics significantly during these seasons.

Eastern and western aquifers at a distance of about 10 km from the river bank show $\delta^{18}O$ values of -7.5‰ and -8.0‰ respectively (sampling profile 10; Table 4.4b). The relatively more depleted values in the western aquifer may due to the recharge of groundwater from the Upper Ganga Canal.

Brijghat: $\delta^{18}O$ values at Brijghat are similar to Balawali, where pre- and post-monsoon data (-6.5‰ and -7.7‰) are relatively depleted than monsoon season (-6.2‰) in the western aquifer. However, the eastern aquifer data show relatively depleted $\delta^{18}O$ during the pre-monsoon season (-9.2‰) as compared to both monsoon and post-monsoon seasons (-6.1‰ and -7.2‰, Table 4.5). The reason for the large difference (about 2.7‰) in $\delta^{18}O$ values of the western and eastern aquifers during pre-monsoon season, may either be due to poor drainage conditions in the western aquifer (leading to evaporation of the groundwater), or mixing of an enriched water source, such as irrigation return flow. If it is due to poor drainage, the $\delta^{18}O$ value should not vary temporally. To confirm this, $\delta^{18}O$ data of the samples collected from the western aquifer at large distances from the river, and are discussed below:

Groundwater samples collected during January 1994 from the western aquifer show $\delta^{18}O$ values of -3.1‰ and -8.4‰ at a distance of 1.0 km and 5.0 km from the river. (sampling profile 18; Table 4.4b). However, the sample collected from the same site at 1.0 km distance from the river during March 1994 show $\delta^{18}O$ value of -8.5‰ (Table 4.4a). This suggests that the groundwater in this area is dynamic and hence poor drainage may not result in $\delta^{18}O$ enrichment in western aquifer. Therefore, $\delta^{18}O$ enrichment may be because of groundwater recharge from irrigation return flow or flushing of the evaporated water from unsaturated zone during irrigation.

Anupshahr: The average $\delta^{18}O$ values in the western and eastern aquifers at Anupshahr are -7.6‰ and -8.3‰, -8.3‰ and -8.2‰ during pre- and post-monsoon seasons. The relative enrichment in $\delta^{18}O$ of the western aquifer during monsoon (-7.4‰) could be due to rainfall recharge coupled with evaporation from groundwater.

Groundwater in Anupshahr area is relatively depleted in $\delta^{18}O$ (-9.8‰; August, 1994) and the sample from Anupshahr branch of Upper Ganga Canal shows a value of -9.7‰ (Table 4.4a). The data indicate that $\delta^{18}O$ of groundwater in the western side of the river is possibly influenced by the canal seepage in Anupshahr area, which in turn contributes to the river.

During January 1994, the $\delta^{18}O$ data of the eastern aquifer along profile no. 21, about 1.0 km (-10.6‰ at Biharipur), 5.0 km (-7.9‰ at Pathra) and 10.0km (-6.6‰ at Gangesri) away from the river, show that the effect of the river diminishes with increasing distance. However, the samples collected from profile no. 23 (Table 4.4b) in the eastern aquifer, at 5.0 km (-7.8‰ at Tomi) and 10.0km (-7.8‰ at Bahat Karan) away from the river indicate that the groundwater is relatively depleted than local rainfall index. Since there is no canal towards the eastern aquifer near Anupshahr, the depletion in groundwater $\delta^{18}O$ may be due to seepage from river Ganga.

Rajghat(Narora): At Rajghat, the groundwater from western aquifer shows an average $\delta^{18}O$ value of -7.3‰ in both pre- and post-monsoon seasons and -6.8‰ in monsoon season. This is similar to that observed at many other sites in the present study. The average $\delta^{18}O$ values of the groundwater from the eastern aquifer is closer to that of the river during all the seasons. (-8.1‰, -8.1‰ and -7.9‰).

4.4.3 Isotopic Characteristics of the River

The $\delta^{18}O$ data of river Ganga have been used to understand the effect of evaporation and also the lateral mixing characteristics, in addition to its interaction with the groundwater.

4.4.3.1 Effect of evaporation on river Ganga

The $\delta^{18}O$ evaporative enrichment in river Ganga is difficult to assess, because the maximum evaporation occurs during summers, while the mixing of relatively heavier groundwater is likely to be a dominant process. However, an attempt has been made to assess the effect of evaporation using Upper Ganga Canal as a proxy. The canal offtakes at Hardwar and therefore its $\delta^{18}O$ value must be similar to that of river Ganga at Hardwar. Since Upper Ganga Canal does not receive any additional inflow along its path, the possibility of mixing with water having different $\delta^{18}O$ value is remote. Therefore, any change in the $\delta^{18}O$ of Upper Ganga Canal must be brought about by evaporation only. Keeping this in view, a sample was collected during May 1995 from the Anupshahr branch of Upper Ganga Canal, and it shows a $\delta^{18}O$ value of -9.7‰. In the same period, river Ganga at Hardwar shows $\delta^{18}O$ content of -11.6‰ (Table 4.4a). Hence, it is inferred that there is an enrichment of about 2‰ in the canal water during its traverse from Hardwar to a point near Anupshahr (about 200 km from Hardwar).

The effect of evaporation in changing the $\delta^{18}O$ content is not observed in all the seasons. The $\delta^{18}O$ and δD data of samples, collected in March 1994 (Table 4.4c) plotted alongwith the New Delhi Meteoric Water Line (Figure 4.4), show that there is no significant enrichment in the river water. Therefore, it may be inferred that the effect of evaporation in the river may be about 2‰ in the selected reach of the river and the evaporation is not a dominant process to change the isotopic composition of the river in all the seasons. Further, in this present investigation, the effect of evaporation is negligible as sampling has been carried out at every 30 to 60 km length of the river course and a moving isotopic index of the river has been considered for each site.

4.4.3.2 Lateral mixing in the river

As Ganga is a large river and there is no major lateral surface inflow to the river, the problem of lateral mixing in altering the isotopic signatures of the river in the selected reach does not exist. However, the inflow of groundwater might modify the isotopic composition of the river. The extent of lateral mixing of the river was therefore assessed in March 1993. Samples

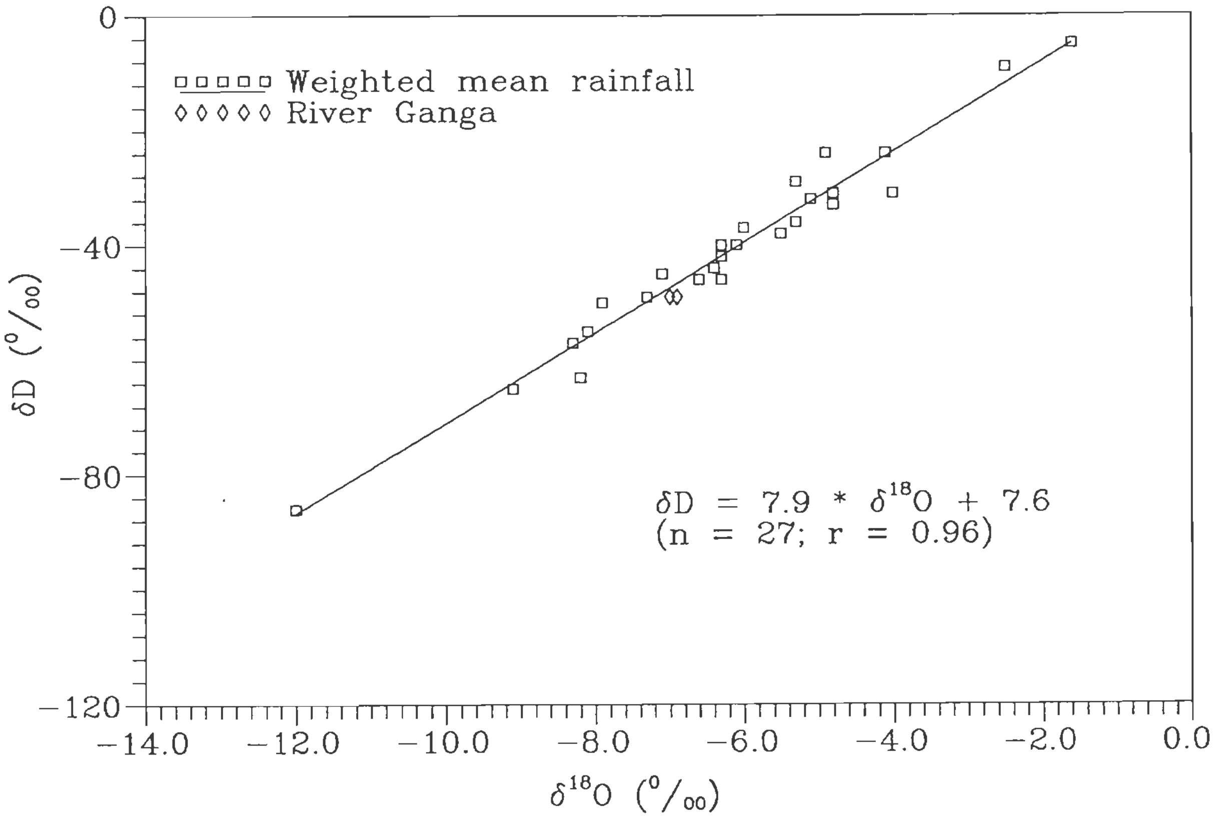


Figure 4.4 Meteoric water line of New Delhi rainfall. Data of river Ganga at Rajghat (March, 1993) are also shown.

collected from lateral extremities of the river at Rajghat (Table 4.4c) show that there is no difference (-6.9‰ and -49‰; -7.0‰ and -49‰) in their isotopic composition, which indicated that the river is laterally well mixed.

4.4.3.3 Spatial variation of $\delta^{18}O$ of river Ganga during different seasons

Pre-monsoon season: It is observed from Figure 4.5a that in river Ganga, during pre-monsoon season, there is a progressive enrichment in $\delta^{18}O$ from Hardwar to Brijghat, and a gradual depletion of $\delta^{18}O$ between Brijghat and Rajghat. At Brijghat, the eastern aquifer is relatively more depleted in $\delta^{18}O$ than the river and western aquifer (Figure 4.5a). It appears that the enrichment in the $\delta^{18}O$ of the river is more due to discharge that occurs from western aquifer than from eastern aquifer. The pattern of $\delta^{18}O$ variation in the river matches more closely with variation observed in the western aquifer than that in the eastern aquifer in all the sites (Table 4.5). Groundwater level and $\delta^{18}O$ variations in both the aquifers observed during pre-monsoon season, indicate that the groundwater contribution seems to have taken place from both the aquifers to the river in the reach from Hardwar to Balawali and only from western aquifer in the reach from Rawalighat to Rajghat.

Monsoon season: During monsoon season (June 1993), as seen from Figure 4.5b(i), the $\delta^{18}O$ of the river varies between -7.8‰ and -8.8‰ at different sites, indicating that there is very little variation in isotopic composition. During August 1993 (Figure 4.5b(ii)), the $\delta^{18}O$ of the river shows continuous enrichment from Hardwar to Rajghat. The enrichment may be attributed to the presence of Madhya Ganga Canal Reservoir, return flow of bank storage and contribution of groundwater. Groundwater levels at Brijghat also indicate effluent conditions (Figure 4.2c). The groundwater closer to the river also show higher $\delta^{18}O$ values (Figure 4.5b(ii)) except at Rawalighat, where it is more depleted than the river. As noted earlier, the enrichment in the aquifers during monsoon season, would have held because of higher recharge due to rainfall, shallow water-table condition followed by evaporation from groundwater. This also indicates that the river from Hardwar to Narora is not influent during the monsoon season.

Post monsoon season: It is observed from Figure 4.5c that the river shows a progressive enrichment between Hardwar (-10.6‰) and Narora (-8.2‰) during the post-monsoon season. Higher $\delta^{18}O$ values at Rawalighat may be attributed to the presence of Madhya Ganga Canal

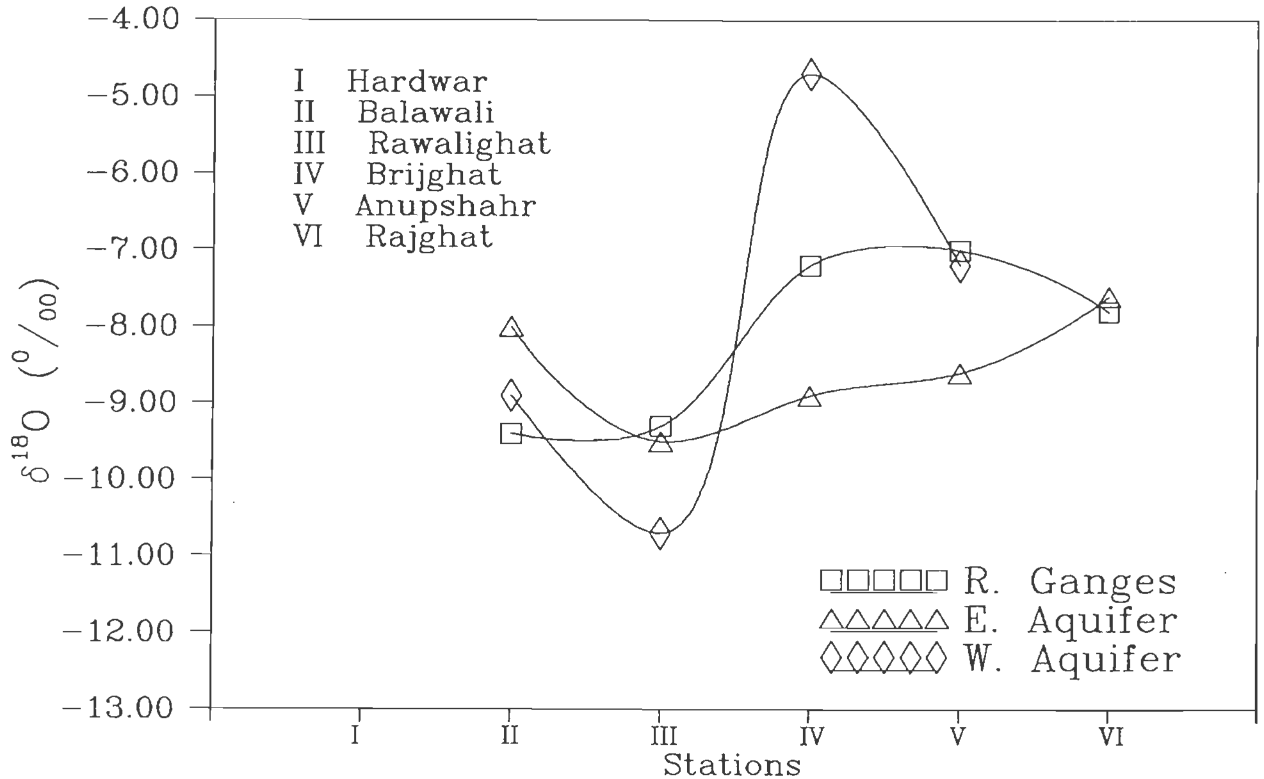


Figure 4.5a Isotopic variations in river Ganga and groundwater during pre-monsoon season (01/93).

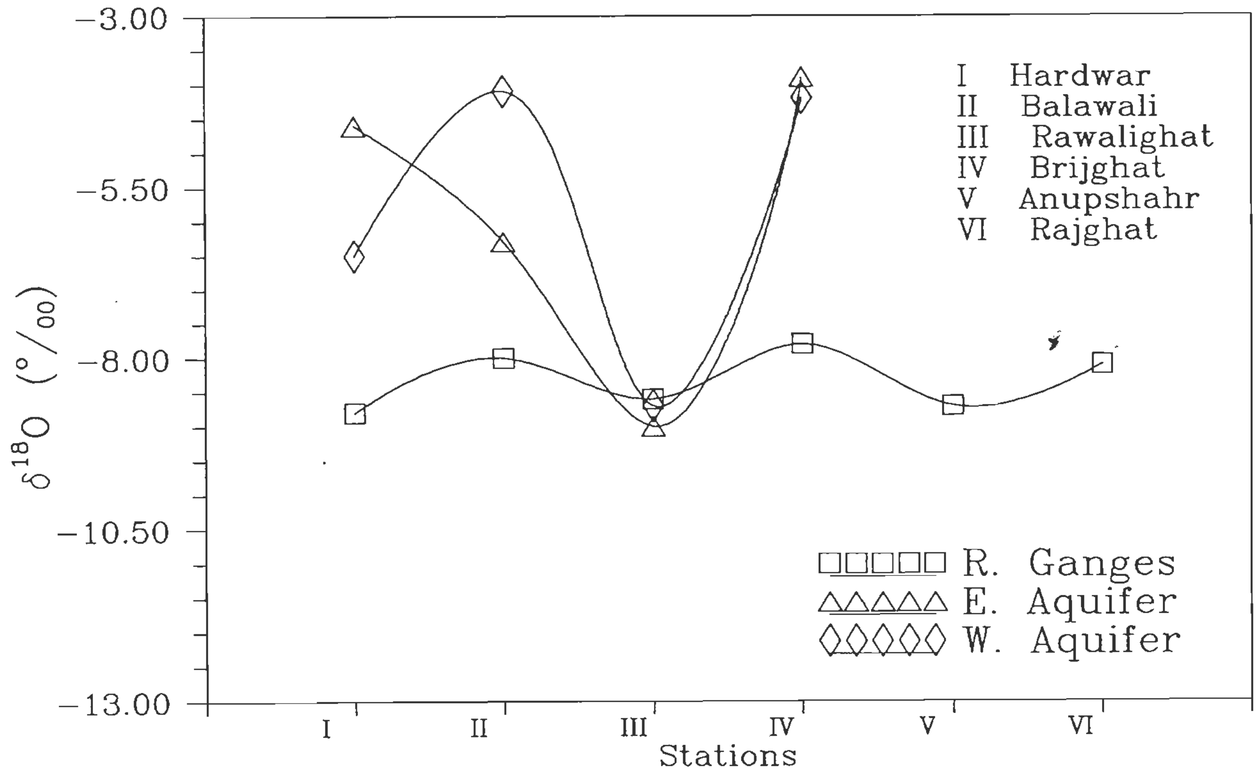


Figure 4.5b(i) Isotopic variations in river Ganga and groundwater during monsoon season (06/93).

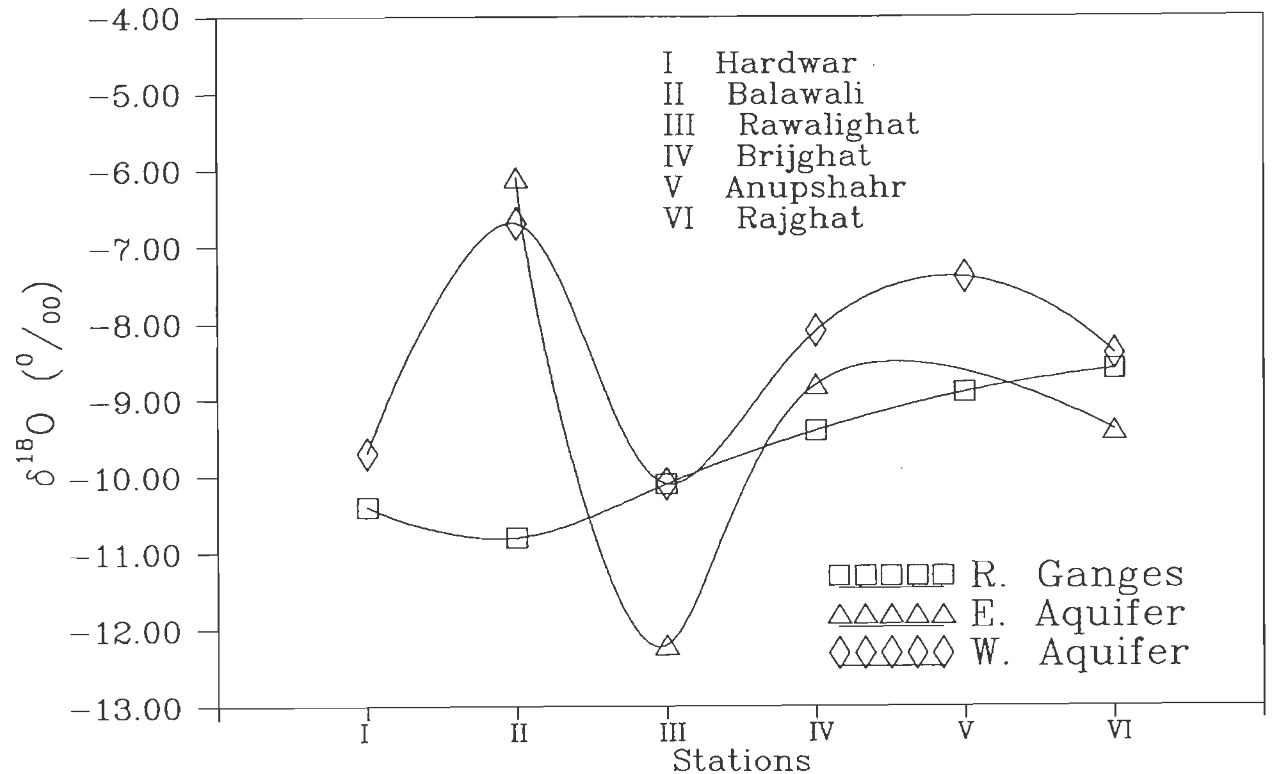


Figure 4.5b(ii) Isotopic variations in river Ganga and groundwater during monsoon season (08/93).

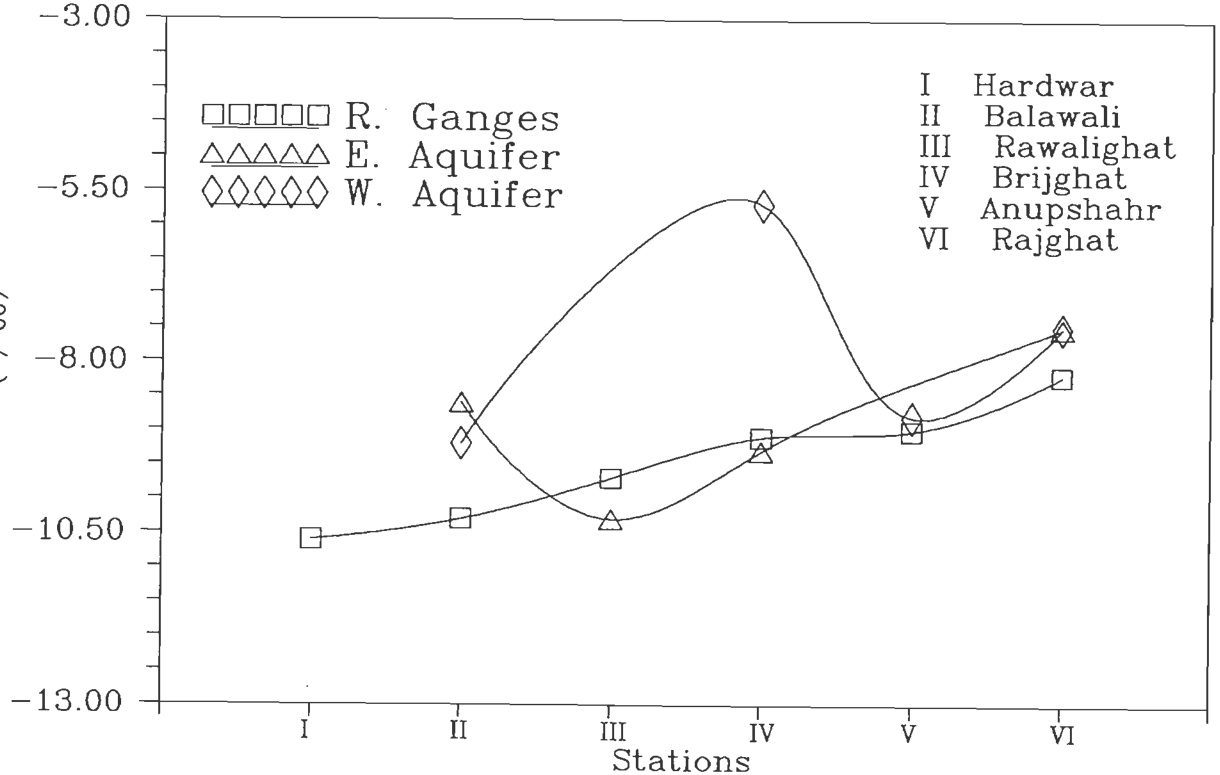


Figure 4.5c Isotopic variations in river Ganga and groundwater during post−monsoon season (11/92).

Reservoir which shows $\delta^{18}O$ of approximately -9.0‰ during monsoon and pre-monsoon seasons. However, the enrichment between Rawalighat (-9.7‰) and Brijghat (-9.1‰) could have occurred due to bank storage and groundwater inflow. Groundwater closer to the river has $\delta^{18}O$ of -5.7‰ and -9.3‰ in the western and eastern aquifers respectively (Table 4.4a). Although, there is little change in the $\delta^{18}O$ of the river between Brijghat and Anupshahr, the flow from groundwater to the river would have continued in addition to the flow from bank storage. It has already been observed that there is an influence of groundwater on the river during monsoon season itself. However, during the post-monsoon season, the effect is not seen in the $\delta^{18}O$ values of river at Anupshahr, where the bank storage has $\delta^{18}O$ of -8.8‰ and the groundwater -9.2‰ to -9.8‰. Mixing of different proportions of bank storage and groundwater would nullify the change in $\delta^{18}O$ value of the river. The enrichment in $\delta^{18}O$ of the river between Anupshahr and Narora may be due to the groundwater contribution.

4.4.4 Estimation of Groundwater Contribution to River Ganga

In this present investigation, the river Ganga between Hardwar and Rajghat (Narora) has been divided into five sub-reaches, in order to estimate groundwater discharge to the river Ganga. The five sub-reaches are: Hardwar to Balawali, Balawali to Rawalighat, Rawalighat to Brijghat, Brijghat to Anupshahr and Anupshahr to Rajghat (Narora). The division is based on the influence of different water bodies such as, Madhya Ganga Canal Reservoir, Ram Ganga Feeder Canal and groundwater inflow on the isotopic characteristics of the river. In order to estimate the groundwater contribution to Ganga river in the study area, a moving isotopic index approach has been adopted as against a fixed index. In the moving index approach, the $\delta^{18}O$ value of the river at the upstream site of the selected sub-reach is considered as the river index. The $\delta^{18}O$ value of groundwater at the downstream end of selected sub-reach and occurring closer to the river has been selected as the groundwater index, provided the groundwater levels are above the river water levels.

Estimation of groundwater discharge to the river in the Hardwar - Balawali, Brighat - Anupshahr and Anupshahr - Rajghat sub-reaches, has been attempted by using two-component mixing model, as given in section 2.6.1. Balawali - Rawalighat sub-reach has been excluded because of the presence of the reservoir. The $\delta^{18}O$ values observed in the western aquifer have been considered for the groundwater index as this aquifer is contributing water to the river. The results are presented in Table 4.6a. The groundwater discharge into the river reach between Hardwar and Balawali has been estimated by two-component mixing model, and the results are presented in Table 4.6b. Further, by back calculating the specific discharge with two- component mixing model results (using the discharge data of the river at Hardwar).The specific discharge of groundwater to river ranges from 13.4 m^2/day during June 1993 to 26.8 m^2/day during March 1994.

A three-component mixing model has been used to account for the influence of Ram Ganga Feeder Canal for estimating groundwater contribution to the river, in the reach between Rawalighat and Brijghat (38).This model has been used to check the assumption whether $\delta^{18}O$ value of the western aquifer observed closer to the river is the appropriate groundwater index. The groundwater contribution to the river has been estimated from the discharge and isotopic data of river Ganga at Rawalighat, Brijghat and Ram Ganga Feeder Canal for January 1994 (Table 4.7). The results were then compared with the results of the following water balance equation by using the river discharge data at Rawalighat and Brijghat:

$$Q_B = Q_R + Q_C + Q_G \qquad (5.5)$$

where, Q is the discharge in m^3/s, subscripts B, R, C and G represent river Ganga at Brijghat and Rawalighat, Ram Ganga Feeder Canal and groundwater respectively.

It is seen from Table 4.7, that the results computed by the three-component model and the water balance method compare well. Therefore, the $\delta^{18}O$ of groundwater from the western aquifer is the true groundwater isotopic index. Since, discharge data of Ram Ganga Feeder Canal were available for the period January 1992 to December 1994, the three- component model has been used in conjunction with Equation 4.5, to back calculate the isotopic index of Ram Ganga Feeder Canal. The computed isotopic values are given in Table 4.7. These values (-6.6‰ to -9.1‰) are well within the range of isotopic variation of the study area. Hence, the results arrived are considered as reliable.

Table 4.6a Results of two-component model for the estimation of groundwater contribution to river Ganga at Balawali, Anupshahr and Rajghat.

Month	Site	$\delta^{18}O_m$	$\delta^{18}O_r$	$\delta^{18}O_g$	% groundwater
November 1992	Balawali	-10.3	-10.6	-9.2	21.4
	Anupshahr	-9.0	-9.1	-8.8	33.3
	Rajghat	-8.2	-9.0	-7.5	53.3
January 1993	Balawali	-9.4	-10.0	-8.9	55.5
	Anupshahr	-7.0	-7.2	-7.2	-
	Rajghat	-7.8	-7.0	-8.3	61.5
February 1993	Balawali	-9.6	-10.3	-8.7	43.8
	Anupshahr	-8.4	-9.0	-8.0	60.0
	Rajghat	-7.4	-8.4	-7.6	-
June 1993	Balawali	-8.0	-8.8	-4.1	17.0
	Anupshahr	-8.7	-7.8	-9.8	45.0
	Rajghat	-8.1	-8.7	-7.5	50.0
August 1993	Balawali	-10.8	-10.4	-6.7	-
	Anupshahr	-8.9	-9.4	-7.4	25.0
	Rajghat	-8.6	-8.9	-8.4	60.0
January 1994	Balawali	-4.6	-8.4	-6.4	-
	Anupshahr	-6.9	-6.3	-9.1	21.4
	Rajghat	-8.3	-6.9	-7.0	-
March 1994	Balawali	-6.7	-10.0	-5.1	67.3
	Anupshahr	-8.4	-6.2	-9.8	61.1
	Rajghat	-7.6	-8.4	-5.7	29.6
May 1995	Balawali	-10.8	-11.6	-10.2	57.1
	Anupshahr	-11.0	-8.1	-7.2	-
	Rajghat	-11.3	-11.0	-8.9	-

Table 4.6b Groundwater contribution to river Ganga, in the reach between Hardwar and Balawali, using isotopic mass balance equation. Q - discharge; B - Balawali and GW - Groundwater.

Month	% Groundwater at Balawali	Q_B* (m^3/s)	Q_{GW}* (m^3/s)	q' (m^2/day)
11/92	21.4	83.4	17.9	19.3
01/93	54.5	36.4	20.1	21.7
02/93	43.8	32.9	14.4	15.6
06/93	17.0	72.9	12.4	13.4
03/94	67.3	36.8	24.8	26.8

* Computed using isotopic mass balance equation and river discharge at Hardwar.

Table 4.7 Results of three-component mixing model for estimating the groundwater component in the river flow at Brijghat (subscripts B, R, C and G represent river Ganga at Brijghat and Rawalighat, Ram Ganga Feeder Canal and groundwater respectively; As the discharge data of river Ganga is classified, the discharge, Q, is given as percentage to the total flow in the river at Brijghat)

Months	Water balance results			Three-component model results				
	Q_R %	Q_C %	Q_G %	$\delta^{18}O_B$ ‰	$\delta^{18}O_R$ ‰	$\delta^{18}O_C$ ‰	$\delta^{18}O_G$ ‰	Q_G %
Nov-92*	-	-	-	-9.1	-9.7	-	-4.7	12.0
Jan-93	26.2	62.6	11.2	-7.2	-9.3	-6.9	-4.7	14.9
Feb-93	25.8	70.1	4.1	-9	-9.5	-9.1	-4.7	4.6
Mar-93	35.4	45.4	18.7	-6.7	-4.5	-9.1	-5.1	19.1
Jun-93*	86.0	-	14.0	-7.8	-8.6	-	-4.2	18.2
Aug-93*	-	-	-	-9.4	-10.1	-	-8.1	35.0
Jan-94	23.1	41.1	35.8	-6.3	-7.7	-6.9	-4.7	35.7
Mar-94	29.1	41.7	29.1	-6.2	-7.8	-6.6	-4.1	30.0
May-95*	-	-		-8.1	-10.9	-	-7.2	75.7

* As there was no discharge from Ram Ganga Feeder Canal, the river at Brijghat has been considered as a two-component mixture.

4.5 Discussions

It has been seen that the results of some studies indicate that river Ganga is effluent during non-monsoon season (145, 105, 146). The results of the present investigation confirm that there is an overall groundwater discharge to the river, but with noticeable temporal variation. This is evident from isotopic characteristics of both river and groundwater.

The specific discharge of groundwater, into the river Ganga in the reach between Hardwar and Balawali, calculated by isotopic technique is much higher than those calculated by Dupuit's method. The discrepancy seems to be due to water levels observed in wells aligned linearly and very close to the river, which probably represent not the true groundwater hydraulic gradient but an apparent gradient as clearly seen from the water level plots. However, the results of the three-component mixing model compare well with those of channel water balance, indicating that the stable isotope technique is a reliable tool to estimate the groundwater component in the river Ganga during different seasons.

The isotopic data of groundwater on either side of the river indicate that the river is effluent on one side, and influent on the other, atleast in selected zones. Earlier studies confirm the existence of such conditions, in selected reaches of the river (145, 146). These conditions may lead to shifting of the river course till equilibrium is attained, where the base level of groundwater table will be determined by the river. The differential behaviour of the river is probably due to neo-tectonic movements, as reported in northern part of the study area (114).

Further, the differential behaviour of river Ganga near Anupshahr area, is also probably related to the geological structures. Several faults in Gangetic basin have been identified (90). The area to the west of Ganga river are relatively elevated as compared to the area lying east of the river. This could lead to groundwater flow from the western aquifer to the river, and river seepage into the eastern aquifer. However, this has to be verified by detailed analyses of the lithologs of the area, as well as the groundwater level data.

CHAPTER 5

CONCLUSIONS AND SCOPE FOR FUTURE STUDIES

5.0 CONCLUSIONS AND SCOPE FOR FUTURE STUDIES

5.1 Conclusions

The surface water and groundwater interaction studies, mainly based on stable isotope characteristics, in two distinct fresh water environments viz. Nainital lake and Ganga river lead to the following conclusions:

Nainital Lake

1. Rainfall data indicate that in Nainital area, about 85% of the annual rainfall is received during monsoon season (June - September). Direct precipitation over Nainital lake accounts for about 10% of annual inflow to the lake.

2. Discharge from some of the upstream springs and some portion of the domestic sewerage (not linked to the town sewerage system) contribute to the flow in the drains, which join the lake. The inflow to the lake through drains accounts for about 10% of the annual inflow to the lake.

3. Surface runoff from the lake catchment contributes to the lake as surface inflow. The surface inflow, as calculated by the Soil Conservation Service - Curve Number (SCS-CN) method is comparable to that computed by using the Lake Level Trend Analysis (LLTA) method developed for Nainital lake. The surface inflow to Nainital lake accounts for about 30% of the annual inflow to the lake.

4. Evaporation losses from Nainital lake, as estimated by the Modified Penman method, indicate that annual losses from the lake due to evaporation account for about 10%.

5. Surface outflow from Nainital lake, as estimated from sluice gate operation records, accounts for about 35% of the total annual outflow from the lake.

6. The results of environmental tracer investigations, indicates that the springs located in Balia ravine (downstream side of Nainital lake) are hydraulically connected to the lake.

This is confirmed by isotopic data obtained from Balia ravine springs. The results obtained by comparing the isotopic data for different seasons indicate that the epilimnion zone of the lake is the main contributing source for the Balia ravine springs.

7. It has been found through isotopic tracer investigations that the proportion of lake water in the Nainital town water supply varies from 25% to 80% during different seasons. The pumping data indicate that the outflow through seepage towards the northern bank of the lake accounts for about 40% of the total annual outflow of the lake.

8. Water balance studies carried out for the total inflow show that the sub-surface contributes 50% of total annual inflow to the lake. The sub-surface outflow is about 55% of the total annual outflow from the lake. It shows that Nainital lake is a 'flow - through' type, with substantial groundwater inflow and lake seepage.

9. Chloride mass balance indicates that annual groundwater inflow to Nainital lake is about 5.99m or 55% of the total annual inflow. Further, the annual seepage from the lake is about 6.78m or about 59% of the total annual outflow from the lake.

10. Isotopic investigations carried out in Nainital lake catchment reveal that the weighted annual mean $\delta^{18}O$ of rainfall is about -11.3‰. The estimated altitude effect on the rainfall in Nainital area in $\delta^{18}O$ is about -0.34‰ and in δD about -2.4‰.

11. A multiple linear regression model has been developed for generating $\delta^{18}O$ data of rainfall in Nainital area, by employing the meteorological parameters such as, monthly mean relative humidity, monthly rainfall and monthly mean air temperature.

12. δ_E was estimated by Craig and Gordon Linear Resistance model. The results of the model indicate that δ_E is controlled mainly by the relative humidity. The estimated δ_E is heavier during months of higher relative humidity, as compared to the months of lower relative humidity.

13. A lower limit for isotopic enrichment of a fresh water lake has been identified by analysing the Craig and Gordon Linear Resistance model. The lower limit is determined by the difference between isotopic composition of the lake (δ_L) and that of the atmospheric vapour above it (δ_a). The evaporative enrichment of the water body will continue if (δ_L - δ_a) is greater than the equilibrium enrichment factor.

14. The temperatures of Nainital lake during December to February indicate that the lake is thermally well-mixed and homogeneous. Further, it is seen that the thermocline, which develops in March disappears totally in November. The temperatures in epilimnion zone show large variations between March and November, while in the hypolimnion zone temperatures vary within a small range, which is identical to that observed in winters.

15. In winters, Nainital lake is isotopically homogeneous and well-mixed. During other seasons, the isotopic characteristics of epilimnion and hypolimnion are different from each other. During summer, the effect of evaporation renders the epilimnion comparatively enriched in $\delta^{18}O$, and during monsoon, surface inflow renders the epilimnion relatively depleted in $\delta^{18}O$.

16. The Isotope mass balance method indicates that groundwater inflow to the lake is about 5.1m or 51% of the total annual inflow, and the seepage from the lake is about 5.88m or 56% of the total annual outflow.

17. The results of both chloride and isotope mass balance methods corroborate the results of water balance method. Water retention time - WRT (volume/outflow) as computed for Nainital lake by isotopic mass balance, chlorine mass balance and conventional water balance methods is about 1.93y, 1.77y and 1.92 y respectively. The WRT will be even lesser for years with higher annual rainfall.

18. Hydrogeological investigations indicate that the shale formation, which occupies about 50% of the lake catchment area, is not a suitable aquifer. However, the catchment area has well developed lineaments and faults. The hydrologic investigations conducted along

these lineaments indicate higher infiltration capacity. Therefore, it is inferred that most of the groundwater inflow to the lake might be occurring along these zones.

19. Seepage from Sukhatal lake appears to be a major recharge source for Nainital lake and any activity in Sukhatal lake may affect the Nainital lake.

20. Analyses of the long-term surface outflow and annual rainfall data, indicate reduction in surface outflow for a given amount of annual rainfall during the past three decades. This reduction is due to the complementary increase in pumping in the northern banks of the lake. It is therefore, inferred that any change in the quantity of pumping may affect the availability of water in Nainital lake.

21. The springs located in Balia ravine are sustained by epilimnion zone of Nainital lake. Therefore, reduction in the lake water level below a depth of about 6 m from surface, may result in the drying up of springs in Balia ravine.

River Ganga Between Hardwar and Narora

1. It has been found from the groundwater level measurements that both western and eastern aquifers contribute to the river flow at Balawali, Brijghat and Anupshahr, in the non-monsoon seasons. Whereas, $\delta^{18}O$ data of groundwater in the eastern aquifer near Anupshahr, shows that the influence of the river decreases with increase in distance. At Rajghat (Narora), it appears from the groundwater level data that the river receives groundwater from the western aquifer, and seeps towards the eastern aquifer during the same period. These indicate that the nature of river and groundwater interaction is different in different reaches of the river, in the study area.

2. The $\delta^{18}O$ variation in the river at Brijghat, Anupshahr and Rajghat, during pre-monsoon seasons is comparable to the variation observed in western aquifer, and therefore, the western aquifer seems to be a major source of water to the river during pre-monsoon period.

3. Analysis of the rainfall isotopic data from New Delhi station shows that the long-term monsoon weighted average for $\delta^{18}O$ is -6.4‰. Therefore, it is likely that the groundwater $\delta^{18}O$ in the study area might be around -6.4‰.

4. The frequency distribution analysis of $\delta^{18}O$ were carried out by using samples pertaining to groundwater collected closer to the river, during pre-monsoon season. The results show the western and eastern aquifers have different peak values viz. -8.5‰ and -9.5‰ respectively. It indicates that different processes are involved in influencing groundwater isotopic characteristics of the two aquifers. Spatial variation in $\delta^{18}O$ values of the aquifers is considerable and it is due to various hydrological processes. The western aquifer is recharged mainly by precipitation, seepage from Upper Ganga Canal, and also by the irrigation return flow. On the other hand, the eastern aquifer is recharged mainly by precipitation, seepage from Eastern Ganga Canal (restricted only to the northern part of the study area), Ram Ganga Feeder Canal and in certain zones by the river Ganga.

5. The frequency distribution analysis of $\delta^{18}O$ were carried out by using samples pertaining to groundwater collected closer to the river, during monsoon and post-seasons. The results show the western and eastern aquifers have different peak $\delta^{18}O$ values of -6.5‰ and -5.5‰ respectively. The enrichment of $\delta^{18}O$ during these seasons reflects the influence of rainfall recharge to the groundwater. The enrichment is also possible due to evaporation of groundwater, as groundwater table closer to river at many sites which are found to be less than one to two metre below the ground level.

6. A plot of $\delta^{18}O$ and δD pertaining to river Ganga and those pertaining to New Delhi rainfall shows that there is no discernible enrichment effect on isotopic characteristic of the river. This signifies that evaporation may not be a major factor for isotopic variations in the river reach between Hardwar and Narora.

7. The mean $\delta^{18}O$ values indicate a progressive enrichment in $\delta^{18}O$ during pre-monsoon season from Hardwar to Brijghat river Ganga, whereas it becomes more negative between Brijghat and Rajghat. It is observed from the isotopic data of both river and aquifer

systems that the enrichment of $\delta^{18}O$ in the river could be due to the contribution of groundwater from western aquifer rather than from eastern aquifer.

8. The specific discharge of groundwater into Ganga river between Hardwar and Balawali has been computed by using two-component mixing model and it ranges from 13.4 m^2/day in June 1993 to 26.8 m^2/day in March 1994. These are found to be higher than those calculated by using Dupuit's method. This discrepancy may be due to linear alignment of the piezometers normal to the flow direction of the river and consequent estimation of the apparent hydraulic gradient of groundwater than true gradient.

9. Groundwater contribution to river Ganga between Rawalighat and Brijghat is estimated by using a three-component mixing model. The results compare well with that of channel water balance method. This substantiates the conjecture that the $\delta^{18}O$ value of groundwater occurring close to the river, particularly when the groundwater level is higher than the river level, is the appropriate isotopic index of groundwater.

10. The contribution of groundwater to the flow in river Ganga at Balawali, Anupshahr and Rajghat during non-monsoon season ranges from 21% to 67% at different sites, while during monsoon season it ranges from 17% to 60%. The quantity of groundwater in the river at Brijghat varies from 5% to 76% during non-monsoon season, and from 18% to 35% during monsoon season. During monsoon season, seepage from river Ganga is retained as bank storage, which subsequently flows back into the river.

This present investigation, on surface water and groundwater interaction in two contrasting environments, viz. a nearly static water body and a flowing water body reveals that different approaches are to be followed for selecting isotopic indices for use in multi component mixing models. Selection of an appropriate isotopic index for a surface water body is governed by its water retention time and the nature of its input sources. In case of a flowing water body such as a river, the moving isotopic index approach should be followed to avoid the effect of evaporation. The effect of evaporation in riverine systems, unlike in lake systems, is not extensively documented. In case of a standing water body such as a lake, the temporal average will be sufficient as the mixing processes and relatively longer water retention times give a characteristic isotopic value.

5.2 Scope for Future Studies

1. In this present investigation, stable isotope technique was applied in two different environments, viz. the Kumaun Lesser Himalaya and the Gangetic plains. Stable isotope data in these regions are few and scattered for hydrological studies. As part of this present investigation, isotopic characterisation of rainfall was carried out in the Kumaun Lesser Himalaya. Isotopic data of precipitation from Parbati Valley, north-west Himalaya, were also collected from literature and compared to those data collected for this present investigation. It was observed that the isotopic characteristics of precipitation, particularly the d-excess parameter, are not found to be comparable. This may be due to different sources of precipitation in different regions of the Himalaya. Therefore, the results of the present investigation may not be applicable to hydrological studies in other parts of the Himalaya. It was further observed, from the isotopic characteristics of rainfall that the altitude effect shows significant variations with different meterological conditions. In view of these, future studies in the Himalayan region may be oriented towards understanding the factors that control isotope ratios in precipitation. As the Himalayan region gives rise to several large river systems, such studies will improve the potential use of isotope tools in the hydrological investigations of these rivers.

2. In the investigations carried out on river Ganga, an attempt was made to use the Upper Ganga Canal as a proxy to study the effect of evaporation on isotopic characteristics of the river. This method could further be explored and developed. Although, the Upper Ganga Canal runs parallel to the main river with similar environmental conditions, influence of groundwater is not observed in the isotopic characteristics of the canal, since the canal is constructed above the ground level. The existence of this set-up may be used to evaluate the extent of evaporative enrichment on canal water, and the findings could be used for not only to evaluate evaporation loss from the river, but also to estimate groundwater contribution to the river more accurately.

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