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STUDY OF LAKE NAINITAL - GROUNDWATER INTERACTION USING ISOTOPE TECHNIQUES

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ABSTRACT

The lake - groundwater interaction study for Lake Nainital, Kumaun Himalaya has been carried out using isotope mass balance method in conjunction with conventional water balance method. Based on the hydrological investigations carried out a conceptual model for the water balance of lake Nainital has been developed. All the components of the lake water balance, except the sub-surface ones, have been measured / estimated using standard methods.

To assess the subsurface outflow from the lake by conventional method, the hydraulic interconnection of the lake and downstream springs were investigated using environmental tracers. Analyses of hydro-chemical data indicate that Lake Nainital and the few downstream springs viz. Sariyatal and Balia ravine springs are hydraulically interconnected and that the other downstream springs such as those located to the west of the lake catchment, and in Kailakhan area are not connected to the lake. In order to complement the information obtained through the hydrochemistry, stable isotopic investigations were carried out. During winter, when the lake is well mixed, the Balia Ravine Springs showed an δ^{18}O value (–9.5‰) close to that of the Lake (-9.6‰), while most of the other springs showed a more depleted value (<-9.6) conforming the results obtained from the hydro-chemical analyses.

In order to assess the proportion of the lake water being tapped in the wells located in the northern bank of the lake, a two-component mixing model was employed using the stable isotope data. The results of the model indicate that the proportion of the lake water in the pumpage varies from 33% to 100% in different seasons.

By combining the discharge data on the spring that are interconnected to the lake and the proportion of the lake water in the pumpage, the total sub-surface outflow from the lake has been estimated (56% of the total outflow). The sub-surface inflow has been estimated as the residual or remainder of the water balance equation, as all other components have been computed / estimated using standard methods. The sub-surface inflow computed by the water balance method is around 49% of the total inflow to the lake. The sub-surface and outflow computed using stable isotope mass balance method are about 51% and 56%, respectively, while that computed using chlorine mass balance method are about 55% and 59%, respectively.

The results show that the estimates of sub-surface inflow to the lake and outflow from the lake obtained through the isotopic and chemical balance compare very well with those obtained through conventional water balance method. The water retention time (τ) of the lake computed using isotopic mass balance approach is about 1.93 years, chemical (chlorine) mass balance is about 1.77 years and conventional water balance is about 1.92 years. The results obtained by all the three methods do not vary significantly from each other (compared to the size of the lake) and compare very well within the error limits. The water retention time computed using isotopic mass balance approach is about 2% lower and that computed using chlorine mass balance results is about 10% lower than the water retention time computed using conventional water balance results. The value that reflects the true water retention time may be the one computed using the isotopic mass balance approach as it was arrived independent of the estimates of pumpage and outflow through springs. Further, isotope mass balance has the advantage over the chlorine mass balance as chlorine may be introduced into the lake and groundwater systems through anthropological activity.

CONTENTS

Preface i

Abstract ii

List of Tables vii

List of Figures ix

1.0 INTRODUCTION 1

2.0 USE OF STABLE ISOTOPES IN SURFACE WATER - GROUNDWATER INTERACTION STUDIES 3

- 2.1 Nomenclature 3
- 2.2 Delta (δ) Notation 4
- 2.3 Isotopic Fractionation 5
  - 2.3.1 Isotopic fractionation by physical processes 5
  - 2.3.2 Isotopic fractionation by chemical processes 5
    - 2.3.2.1 Equilibrium effects 6
    - 2.3.2.2 Non-Equilibrium (kinetic) effects 8
- 2.4 Isotopic Composition in Precipitation (δ_p) 9
  - 2.4.1 Latitude effect 11
  - 2.4.2 Continental effect 11
  - 2.4.3 Seasonal effect 12
  - 2.4.4 The amount effect 12
  - 2.4.5 Temperature relationship 13
  - 2.4.6 Altitude effect 13
  - 2.4.7 Relation between δ_p and geographical/climatological parameters . . 14
  - 2.4.8 $\delta^{18}O$ - δD relationships 14
  - 2.4.9 'd' - excess parameter 16
- 2.5 Lake - Groundwater Interaction Studies 16
  - 2.5.1 Lake evaporation 17
    - 2.5.1.1 Dissolved salt effect 21
  - 2.5.2 Quantitative evaluation of the lake - aquifer interaction 22
    - 2.5.2.1 Well mixed lakes 22
    - 2.5.2.2 Poorly mixed lakes 29
  - 2.5.3 Qualitative studies on lake - aquifer interaction 30

3.0 CHARACTERISTICS OF LAKE NAINITAL BASIN 32

- 3.1 Geology 32
  - 3.1.1 Lithostratigraphy 32
  - 3.1.2 Structure 35
  - 3.1.3 Geochemistry of the rock formations of the Lake basin 36

3.2 Precipitation . . . 38
- 3.2.1 Long-term average precipitation rates . . . 38
- 3.2.2 Seasonal variability . . . 39
- 3.2.3 Orographic effects on precipitation . . . 40
- 3.2.4 Estimation of areal precipitation . . . 41
- 3.2.5 Precipitation quality . . . 42
- 3.2.6 Stable isotope ratios . . . 43
  - 3.2.6.1 Materials and methods . . . 43
  - 3.2.6.2 Local meteoric water line (LMWL) . . . 44
  - 3.2.6.3 'd' Excess Parameter . . . 45
  - 3.2.6.4 Altitude effect in stable isotope ratios . . . 46
  - 3.2.6.5 Local meteorological conditions and δp . . . 47
  - 3.2.6.6 Interannual variability in isotope ratios: . . . 48

3.3 Hydrology . . . 49
- 3.3.1 Landuse Pattern . . . 49
- 3.3.2 Soil properties . . . 51
- 3.3.3 Infiltration . . . 53
- 3.3.4 Evapotranspiration . . . 54
  - 3.3.4.1 Potential Evapotranspiration . . . 54
  - 3.3.4.2 Actual Evapotranspiration . . . 55
- 3.3.5 Drainage . . . 56
  - 3.3.5.1 Flow through the drains . . . 57
- 3.3.6 Chemistry of drain waters . . . 57
- 3.3.7 Stable isotope ratios . . . 59
- 3.3.8 Surface runoff . . . 60
  - 3.3.8.1 Lake Level Trend Analysis Method . . . 60
  - 3.3.8.2 Soil Conservation Service - Curve Number Method . . . 62

3.4 Lake Characteristics . . . 65
- 3.4.1 Morphology . . . 65
- 3.4.2 Estimation of volume . . . 66
  - 3.4.2.1 Change in storage . . . 70
- 3.4.3 Thermal regime . . . 70
  - 3.4.3.1 Vertical exchange coefficient . . . 72
- 3.4.4 Surface Outflow . . . 74
- 3.4.5 Chemistry of lake water . . . 75
  - 3.4.5.1 Materials and methods . . . 76
  - 3.4.5.2 Analyses of data and accuracy . . . 78
  - 3.4.5.3 Characterisation . . . 79
  - 3.4.5.4 Stable isotope ratios . . . 85
- 3.4.6 Lake evaporation . . . 86

3.5 Hydrogeology . . . 89
- 3.5.1 Aquifer properties . . . 92
- 3.5.2 Springs . . . 95
- 3.5.3 Groundwater withdrawal . . . 98
- 3.5.4 Chemistry of groundwater . . . 100
- 3.5.5 Stable isotope ratios . . . 110

4.0 LAKE NAINITAL - GROUNDWATER INTERACTION 114

4.1 Conceptual Model for the Lake Water Balance 114
4.2 Interconnection of Lake Nainital and Downstream Springs 117
4.3 Two-Component Mixing Model for the Estimation of Proportion of Lake Water in the Pumpage (Wo) 118
4.4 Estimation of Sub-Surface Outflow Through Springs 119
4.5 Uncertainties in the Estimation of Water Balance Components 120
4.6 Water balance for the Lake Nainital 121
4.7 Isotope Mass Balance 121
4.8 Chlorine Mass Balance 128

5.0 CONCLUSIONS 130

References 131

List of Tables

3.1. Lithostratigraphic succession of the Mussoorie Group in Nainital Hills of the Krol Belt, Kumaun Lesser Himalayas (Valdiya, 1988) 32
3.2. Chemical composition of dolomitic rocks of Lake Nainital basin 36
3.3. Theissen Weights of different raingauge stations in Lake Nainital Basin. 40
3.4 The isotopic composition of precipitation samples collected at different altitudes in the Nainital basin and downstream area 44
3.5. Estimated altitude effects (per 100 m) in $\delta^{18}O$ and δD for the monsoon precipitation of 1995 at Nainital, Kumaun Himalayas. 45
3.6. Isotopic composition precipitation and the monthly meteorological data of the corresponding month of sampling at the lake site. 47
3.7 Landuse data of Lake Nainital basin using remote sensing data 48
3.8 Landuse data of Lake Nainital basin using conventional approach 49
3.9 Moisture retention characteristics of soil samples collected from different sites in Lake Nainital catchment. 51
3.11 Fluvial morphometric parameters of Lake Nainital basin (After Rawat, 1988) 55
3.12 Details of samples collected from drains 56
3.13. Chemical characteristics of drain waters during non-monsoon season 57
3.14. Chemical characteristics of drain waters during monsoon season 57
3.15. Isotopic characteristics of drain waters 58
3.16. Hydrologic soil classification used in SCS-CN method. 62
3.17. Determination of antecedent moisture conditions based on antecedent rainfall 62
3.18. Determination of composite Curve Number for AMC II based on landuse data of Lake Nainital basin 63
3.19 Surface area of Lake Nainital at different depths of water level. 66
3.20 Sub-basin wise volume of Lake Nainital at different water levels 67
3.22. Vertical exchange coefficients estimated from lake temperature data 72
3.23 Details of samples collected from Lake Nainital 74
3.24 Thermal characteristics of Lake Nainital and Hydometeorologic characteristics of the lake basin during different seasons 77
3.25a Average values of Electrical Conductance, pH and T. Hardness during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital 78
3.25b Average values of Alkalinity, Sulphate and Chloride during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital 79
3.25c Average values of Phosphorous, Nitrogen and Sodium during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital 80
3.25d Average values of Potassium, Calcium and Magnesium during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital 81
3.26 Mean observed $\delta^{18}O$ and δD in the epilimnion and hypolimnion zones of Lake Nainital during different sampling campaigns 82
3.27 Values of measured and estimated parameters used in the estimation of evaporation from Lake Nainital by Penman Method for the year 1995. 85
3.28 Classification of upstream springs based on average discharge and coefficient of variation in the discharge. 92

3.29 Classification of downstream springs based on average discharge and coefficient of variation in the discharge. 94
3.30 Discharge capacities of wells located in the bank of Lake Nainital 96
3.31 Quantity of water pumped (m^3) out from the banks of Lake Nainital in different months during the study period. 97
3.32a Chemical characteristics of Pardhadhara Spring 98
3.32b Chemical characteristics of Lakeview Spring 99
3.32c Chemical characteristics of Doctor House / Chunadhara Springs 99
3.32d Chemical characteristics of Sariyatal Spring 100
3.32e Chemical characteristics of Springs located in Kailakhan area 100
3.32f Chemical characteristics of Sipahidhara Spring 101
3.32g Chemical characteristics of Gupha Mahadev Temple Spring 101
3.33 Comparison of average Total Hardness and CMR along with Coefficients of variation for different springs 102
3.34 Saturation indices of Calcite and Dolomite of the spring waters sampled during different seasons in 1995. 105
3.35 Observed $\delta^{18}O$ values of the upstream springs during different seasons. 107
3.36 $\delta^{18}O$ values of downstream springs. Samples collected during December, 1994. . . 109
3.37 $\delta^{18}O$ values of Gupha Mahadev Temple spring. Samples collected during different seasons. 110
4.1 Results of two-component mixing model using stable isotope data to estimate the proportion of lake water in the well waters being pumped 115
4.2 Results of two-component mixing model using water chemistry data to estimate the proportion of lake water in the well waters being pumped (Wo) 116
4.3 Comparison of the discharge of Sipahidhara spring and that of all springs 117
4.4: Errors due to time averaging of various primary input data to evaporation calculations and the propagated error in the evaporation estimate 121
4.5 Estimated flow components with uncertainties for Lake Nainital 123
4.6: SSi and SSo estimated using isotopic and chemical mass balance approaches 129

List of Figures

3.1 Geological map of Nainital basin showing locations of springs and wells 35a
3.2 Location map of the study area showing Nainital lake and sampling sites 38a
3.3 Correlogram of annual rainfall in Lake Nainital catchment 39a
3.4 Variation of monsoon and non-monsoon rainfall at Manora (Nainital) 40a
3.5 Orographic effect in Rainfall - Lake Nainital basin (1995) 41a
3.6 Variation of 30 year average monthly mean relative humidity at Manora 41b
3.7a Altitude effect in del O-18 during the monsoon season of 1995 based on the rainfall weighted averages 46a
3.7b Altitude effect in del D during the monsoon season of 1995 based on the rainfall weighted averages 46b
3.8 Variation of observed and computed soil moisture in Lake Nainital catchment 51a
3.9 Variation of Potential evapotranspiration (penman method) for Lake Nainital basin. Data pertaining to the year 1995 55a
3.10 Variation of Actual evapotranspiration (soil moisture retention method) for Lake Nainital basin. Data pertaining to the year 1995 55b
3.11 Variation of the lake level with daily rainfall during 1994 61a
3.12 Variation of the lake level with daily rainfall during 1995 61b
3.13a Comparison of daily variation in surface inflow to the lake in 1995 estimated by LTTA and SCS-CN methods 61c
3.13b Monthly variation in the surface inflow to the lake in 1995 estimated by LLTA and SCS-CN methods 61d
3.14a Monsoon season rainfall-runoff relationship in Lake Nainital catchment in 1995 by LLTA method 61e
3.14b Non-monsoon season rainfall-runoff relationship in Lake Nainital catchment in 1995 by LLTA method 61f
3.15 Bathymetric map of Nainital lake (After Hashimi et al., 1993) 65a
3.16 Variation of volume of Nainital lake with depth 67a
3.17 Variation of surface area of Nainital lake with depth 67b
3.18 Temperature variation in Mallital sub-basin of Nainital lake during 1994-95 72a
3.19 Temperature variation in Tallital sub-basin of Nainital lake during 1994-95 72b
3.20 Geological cross-section along Copacabana and St. Loe in the Sher-ka-Danda hill, Nainital. Data source:Ashraf, 1978 90a
3.21 Plot of h/ho versus time for estimation of K in Manora Shales by slug-test using Hvorslev method 93a
3.22 Well hydrograph of DH#2 located in Manora Shales below the spring near Spring Cottage, Nainital 94a
3.23 Well hydrograph of DH#4 located in Manora Shales above the spring near Spring Cottage, Nainital 94b
4.1 Conceptual model showing various water balance components of Nainital lake 114a
4.2 Sulfate-Chloride cross plot for different water sources 117a
4.3 Variation of total cations (TZ^+) of different water sources with time 117b
4.4 Temporal variation in $\delta^{18}O$ of Gupha Mahadev Spring and Lake epilimnion zone 118a
4.5 Propagated error in δ_E in CGLR model results due to errors in input parameters at different values of h 127a

1.0 INTRODUCTION

The interaction between surface water and groundwater is a part of the hydrological cycle. Flow of groundwater to support the surface water and the flow from the surface water body to groundwater are the two main aspects of the interaction process (Wright, 1980). The nature of interaction depends on the nature of the water body whether it is a flowing body such as a river / stream or standing water body such as a lake or reservoir. The size of the water body also determines the effect of interaction as in case of smaller lakes or rivers the quantum of groundwater inflow will have a profound impact on their quality. On the other hand, in case of larger lakes or rivers it is the relative proportion of the groundwater inflow that will effect a change in the water quality.

Groundwater has been recognized as not only a source of the base flow in the streams but also an important contributor during storm flow (Freeze, 1974; Sklash and Fervolden, 1979). This has led the hydrologists in recent years to focus on the physical mechanism of runoff generation (Eshleman et al. 1994; Kendall et al., 1995). However, it is the quantum of interaction that is very crucial for the water resources managers, as the water availability in river / lake has a direct bearing on the irrigation and power projects apart from the need to meet the domestic demands.

There are several conventional methods to assess the river - groundwater interaction such as analyses of river hydrograph and/or groundwater table fluctuations and use of mathematical models. While the conventional methods are reliable the need for extensive periodical measurements of various hydrological parameters makes the methods more laborious and error prone. On the other hand, isotope techniques are precise but can not act as an independent hydrologic tool. However, if the isotope techniques are used in conjunction with the conventional techniques, a more reliable and accurate modeling of the hydrological processes would be possible (Payne, 1981). The isotope techniques are highly useful in defining the problem in hydrological terms particularly to test the hypotheses developed using other techniques in any hydrogeological setting. The basis for use of isotope techniques for surface water - groundwater interaction is that the surface water has a stable isotopic composition that is distinct from the local groundwater. The distinct isotope signature in case of a lake is brought about by evaporative enrichment in the lake. The isotope techniques are helpful in developing / substantiating conceptual models (Bills and Hjalmarson, 1990). The technique is also useful in testing hypotheses such as interconnection between lake and groundwater (Payne, 1970) and in studying the groundwater component in the lake water balance (Dinτer, 1968; Krabbenhoft et al., 1990; La Baugh et al., 1997).

In India practical application of stable isotopes for fresh water hydrological problems are few. Except for some case studies in the arid and semi-arid zones such as investigations for the groundwater recharge source identification, the technique has not been tried extensively. The major reason is non-availability of primary hydrological data for area of interest. To examine the possibility to use the stable isotope tools in surface water - groundwater interaction studies, a high altitude Himalayan lake has been investigated in the present study. The study area was selected on the basis of the fact that the Lake Nainital is a very important source of water for domestic use of the local population.

Lake Nainital is located in the Lower Himalayan range, within a highly disturbed geological setting. The lake is the main source of domestic water supply to the local population living in and around Nainital town. As the town and the lake are a major tourist attraction in northern India, the demand for water for domestic use is increasing with time. It is also important to conserve the aesthetic values of the lake. As no other source of drinking water exist in the area, one of the major concerns is the dependability of the lake as a drinking water source. What are the chances that the lake will get dried up? What are the main recharges sources for the lake? Whether there is any leakage from the lake, if so what is the quantum of leakage? If the surface outflow of the lake diminishes, then it may seriously impair the water quality, deteriorating the already eurtrophied lake.

Much of these concerns can be answered by having the knowledge of the water balance of the lake. Therefore an attempt has been made to compute the lake water balance. Isotope techniques have been used to compute the water balance of a lake, for the first time in India. In using the isotope techniques for lake water balance study the results are arrived independent of information on the subsurface outflow and hence has an advantage over the conventional method.

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

Environmental isotopes are increasingly being used as tools to study various processes in hydrology. Payne (1983) defined environmental isotopes as "those isotopes, both stable and radioactive, which occur in the environment in varying concentrations over which the investigator has no control". This classic definition clearly brings out all the aspects of environmental isotopes. Of the environmental stable isotopes, oxygen and hydrogen find wider application in hydrology because, as part of the water molecule itself, they act as natural tracers.

2.1 Nomenclature:

There are six isotopes of oxygen viz., ^{14}O, ^{15}O, ^{16}O, ^{17}O, ^{18}O and ^{19}O. Of these six isotopes, ^{14}O, ^{15}O and ^{19}O are radioactive with very less half-life of 72 sec., 2 min. and 29 sec. respectively and are not useful for any meaningful hydrological study. The mass abundance of the stable oxygen isotopes are given below:

Isotope	Mass	Abundance %
^{16}O	15.994915	99.759
^{17}O	16.999133	0.037
^{18}O	17.999160	0.204

It is readily seen from the above table that among the heavier isotopes, ^{18}O is around 5.5 times more abundant than ^{17}O. In terrestrial materials the geochemistry of ^{17}O is similar to the more abundant ^{18}O (Gat, 1981). therefore the abundance ratio of $^{18}O/^{16}O$ is considered important in most hydrological studies which involve stable isotopes.

Compared to the six numbers of isotopes in oxygen, hydrogen has only three viz., ^{1}H, ^{2}H and ^{3}H. ^{1}H which is the most abundant hydrogen isotope is also known as Protium and is usually denoted as H. ^{2}H is known as Deuterium and denoted as D. ^{3}H is known as Tritium and denoted as T. Of the three hydrogen isotopes, T is radioactive with a half-life of 12.43 years. The abundance of stable isotopes H and D are given below:

Isotope	Mass	Abundance %
H	1.007825	99.9855
D	2.014102	0.0145

With these isotopes of oxygen and hydrogen several different water molecules are possible, of which the isotopic species $H_2{}^{16}O$, $H_2{}^{18}O$ and $HD^{16}O$ are generally considered in isotope hydrology. The common isotopic water species $H_2{}^{16}O$ has a vapour pressure, slightly different from that of the rarer isotopic water species such as $H_2{}^{18}O$ and $HD^{16}O$.

2.2 Delta (δ) Notation

The stable isotopic ratios are generally expressed as deviation with respect to a standard. The universally adopted standard for oxygen and hydrogen isotopes in water samples is V-SMOW (Vienna - Standard Mean Oceanic Water). The results are denoted by δ, expressed in permil (‰).

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

R stands for the ratio of rarer to the common isotope, subscript x denotes $^{18}O/^{16}O$ or D/H. The advantages in adopting such a convention are many, such as intercomparison of results and better resolution over the absolute ratios. The δ values are also additive, thereby aiding in use of component mixing models. Generally while reporting hydrological studies, the subscripts such as p, a, l, g, i and o are added to δ, which denote precipitation, atmospheric moisture, lake, groundwater, inflow and outflow respectively.

2.3 Isotopic Fractionation

The relative ratio of rarer isotope to the common isotope is of greater relevance in many hydrological studies. The variation in the global distribution of different isotopes in different hydrologic systems are 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.

2.3.1 Isotopic fractionation by physical processes

As noted above, there are different isotopic species of water ranging from the "lighter" $H_2{}^{16}O$ to the heavier $H_2{}^{18}O$ or $HD^{16}O$. The molecular diffusivities of vapour of different isotopic species of water through air has been studied by several investigators (Merlivat, 1978) and has been found that the ratio of diffusivities of lighter molecule to that of the heavier molecule is not same for different isotopic species, even though the difference in the mass of the heavier molecules considered, are not different. The ratio of diffusivities in air is also practically temperature invariant (Gat, 1981). While in the liquid, the molecular self-diffusion coefficients of isotopic water species of equal mass are not equal. As shown by Wang et al. (1953), the intermolecular binding forces are activated by the diffusion process in the liquid state and the effect is more pronounced in case of the hydrogen isotopes than in oxygen isotopes. Other properties such as viscosity is also similarly affected by intermolecular forces, especially by hydrogen bonding and not just by mass difference.

2.3.2 Isotopic fractionation by 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.

2.3.2.1 Equilibrium effects

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 latter. This is because of the differences in the hydrogen bond energy in the liquid phase between the isotopic molecules. This results in the differences in the vapour pressure of the isotopic species of water.

If, the isotopes are randomly distributed in all positions in liquid phase and vapour phase, equilibrium fractionation factor (α^+) is related to the equilibrium constant K, such that

$$\alpha^+ = K^{1/n}$$

where, n is the number of atoms exchanged. For monatomic reactions α = K. For the oxygen isotope exchange between the liquid and vapour phase which are in equilibrium, the following equation holds:

$$H_2{}^{16}O_{(l)} + H_2{}^{18}O_{(v)} \rightleftharpoons H_2{}^{16}O_{(v)} + H_2{}^{18}O_{(l)}$$

Then, equilibrium constant

$$K = [H_2{}^{16}O]_{(v)}\,[H_2{}^{18}O]_{(l)} / [H_2{}^{16}O]_{(l)}\,[H_2{}^{18}O]_{(v)}$$

where [ ] refer to activity coefficients. The above equation may be re-written as:

$$K = ({}^{18}O/{}^{16}O)_l / ({}^{18}O/{}^{16}O)_v = \alpha^+$$

The difference in the ratio of heavier to lighter isotopes in one phase to that of the other is then, is defined by equilibrium fractionation factor α^{+}:

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

where R is the ratio of concentrations of heavier to lighter isotopes.

The equilibrium fractionation factor has been determined by several investigators through vapour pressure measurement of pure isotopic species (Szapiro & Steckel, 1967), through isotopic analyses of water and water vapour in equilibrium (Majoube, 1971; Bottinga and Craig, 1969; and Kakiuchi and Matsuo, 1979), or by dynamic distillation methods (Borowitz, 1962). However, the results presented by Majoube (1971) is widely used in isotope hydrology studies. Majoube (1971) proposed the equation of the following type for α^{+}:

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

Where T is the absolute air temperature in Kelvin, and A, B, and C are coefficients. The values of the coefficients proposed by Majoube (1971) are:

Isotope	A	B	C
$^{18}O/^{16}O$	1137	-0.4156	-0.00207
D/H	24844	-76.248	+0.05261

Evaporation and condensation processes are of special interest in stable isotope geochemical applications in hydrology. 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'. Therefore,

$$\alpha^{*} = \frac{(R_x)_{vapour}}{(R_x)_{liquid}} \tag{2.3}$$

The value of α^* at 0°C is equal to 0.9884 and at 25°C is equal to 0.9907. 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 defined as (Gonfiantini, 1986)

$$\epsilon^* = (1 - \alpha^*)\ 1000\ ‰ \tag{2.4}$$

2.3.2.2 Non-Equilibrium (kinetic) effects

Under natural conditions, the actual isotopic composition of water vapour is significantly more depleted than the values predicted using equilibrium enrichment factors. In other words, the heavier isotopic enrichment in liquid phase is larger than the values computed from the vapour pressure differences of the isotopic species (Dansgaard, 1961; Ehhalt and Knott, 1965; and Craig et al., 1963). The difference between the total enrichment factor (ϵ) and the equilibrium enrichment factor (ϵ^*) is called as excess separation factor or more commonly, kinetic enrichment factor ($\Delta\epsilon$). The kinetic enrichment factor is defined by Craig and Gordon (1965) as follows:

$$\Delta\epsilon = (1 - h) \cdot \left(\frac{\rho_i}{\rho} - 1 \right) \tag{2.5}$$

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. For oxygen isotopes the term (ρ_i / ρ) is close to $(D/D_i)^n$, where, D and D_i are the molecular diffusion coefficients and n is the turbulence parameter. The value of n may range between 0 and 1, bur under most natural conditions a value of 0.5 is more appropriate (Gonfiantini, 1986).

The ratio D/D_i may be calculated using the following expression, obtained through kinetic theory of gas of low concentration diffusion in other gases (Craig and Gordon, 1965):

$$\frac{D}{D_i} = \left(\frac{M_i\ (M + 29)}{M\ (M_i + 29)} \right)^{0.5} \tag{2.6}$$

where, M and M_i are molecular weights of lighter and heavier molecules and 29 is the mean molecular weight of air. Attempts to evaluate the kinetic enrichment factor has also been made Gat (1970) thorough field evaporation pan experiments, by Merlivat (1978) through laboratory experiments and by Vogt (1976) through wind tunnel experiments. Gonfiantini (1986) suggested that since the values reported for $\Delta\epsilon$, by Merlivat (1978) and Vogt (1976) are in very good agreement, the mean of the values derived by the two investigators could be used to adequately evaluate $\Delta\epsilon$, for natural evaporation conditions. He proposed:

$$\Delta\epsilon_{^{18}O}\ ‰ = 14.2(1-h) \qquad (2.7a)$$

$$\Delta\epsilon_{D}\ ‰ = 12.5(1-h) \qquad (2.7b)$$

Zimmermann and Ehhalt (1970), Zimmermann (1979) and a few other investigators made use of deuterium isotopes in lake studies. They contended that in case of deuterium the kinetic enrichment effect is small and that its facilitates in avoiding complications that arise due to estimation of mean relative humidity values. However, Zuber (1983) showed that the kinetic effects in case of deuterium differs very much in field conditions from that in laboratory controlled experiments. His findings confirmed the earlier reports of Gat (1970) on the basis of evaporation pan experiments in lake Tiberias study.

The fractionation factors and enrichment factors give rise to the variation in the isotopic composition of the precipitation and in water bodies subjected to evaporation. The resultant variation of isotopic composition in hydrological systems have been utilised to solve many a hydrological problems.

2.4 Isotopic Composition in Precipitation (δ_p)

It is a well known fact that 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 (Dansgaard, 1953, 1954; Epstein and

Mayeda, 1953; and Friedman, 1953) 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 by Craig (1961). Dansgaard (1964) in his classical review of isotopic data collected from International Atomic Energy Agency / World Meterological Organisation network stations brought out several salient features of the isotopic characteristics in precipitation, relating the distribution pattern to different geographical and environmental parameters such as latitude, altitude, distance from the coast, amount of precipitation and surface air temperature. He also attempted to explain the observed isotopic variations in the atmospheric waters through open system model based on equations that describe separation of gases during distillation, which in turn were proposed by Lord Raleigh (1896).

The Raleigh type model describes the isotopic exchange between the vapour and liquid phases, which are in equilibrium. If the condensation of the vapour proceeds as a slow process with immediate removal of the condensate from the vapour, the isotopic ratio of the liquid and vapour phase will change as:

$$\delta_c = \left(\frac{\alpha_t}{\alpha_0}\right) f_v^{\alpha_m -1} - 1 \tag{2.8a}$$

$$\delta_v = \left(\frac{1}{\alpha_0}\right) f_v^{\alpha_m \ 1} - 1 \tag{2.8b}$$

where, α_t, α_0 and α_m are the fractionation factors at instantaneous temperature t, at initial temperature t_0, and at the mean temperature t_m ($t_m = (t + t_0) / 2$), f_v is the fraction of vapour remaining and δ_c and δ_v are the isotopic ratios of the condensate (liquid or solid) and the vapour phase, respectively. It is obvious that, the Raleigh process then leads to much higher fractionation than under simple equilibrium exchange process. The Raleigh equation for a condensation process is (Dansgaard, 1964):

$$R_v = R_{v_0} \, f^{(\alpha' \ \ 1)} \tag{2.9}$$

where R_v and R_{v0} are the isotopic ratio of the instantaneous and the initial vapour, f is the fraction of residual vapour and α^+ is the fractionation factor. According the above open system model adopted by Dansgaard (1964), the condensate is enriched in the heavier isotopes in comparison to the vapour phase, and as condensation proceeds it will result in more depleted condensates. However, studies on individual storm events revealed that isotopic ratios of successive portions of precipitation may not be monotonously decreasing (Rindsberger et al., 1990). In general, the isotopic composition of precipitation for a given storm depends strongly on the meteorological history of air in which the precipitation is produced and through which it falls (Rozanski et al., 1993).

Dansgaard (1964) also noted that during the fall from the cloud base to the ground, the rain drops are subjected to evaporation and exchange with the environmental vapour. These processes proceed under non-equilibrium conditions (Ehhalt et al., 1963). This process is also called as secondary evaporation. Based on the IAEA/WMO database (presently the programme of collection of isotope data of meteoric waters from stations world over is known as Global Network of Isotope in Precipitation) on monthly composite samples of precipitation the following characteristics were identified (Dansgaard, 1964; Yurtsever & Gat, 1981; Rozanski et al., 1993).

2.4.1 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. The role of temperature is also noticeable in the latitude effect. The comparatively lower temperature condition over mid and high latitude regions results in reduced total precipitable water in the atmosphere. This factor is particularly more pronounced in winter.

2.4.2 Continental effect

The heavier isotope ratio decreases inland from the coast. 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, Amazon basin and Northern India reveal that substantial amount of evapo-transpirated water from the plant covers returns to the atmosphere. This evapotranspiration process reduces the inland gradient of isotopic ratio during certain seasons (Gat and Matsui, 1991). The continental effect in the monsoon precipitation over northern India was studied by Krishnamurthy and Bhattacharya (1991) along the stretch between Calcutta and New Delhi using the groundwater isotopic ratios as proxy. They reported an apparent depletion in the $\delta^{18}O$ to be of the order of -2‰/1000 km. This value is not much different from the results reported by Piexoto and Oort (1983) for precipitation in the European continent, which was -8‰/4500 km.

2.4.3 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. Rozanski et al. (1993) suggested that the seasonal variations could be brought about by a) seasonal variation in total precipitable water, which is more so in mid and high latitude regions, b) seasonally modulated evapotranspiration over continental regions and c) seasonally changing source areas of the vapour and different storm trajectories. The amplitude of seasonal variation of the isotopic ratio also increases with increasing distance from the coast.

2.4.4 The amount effect

Dansgaard (1964) observed an apparent relationship between the amount of rainfall and its isotopic ratio i.e. for greater amount of rainfall, the isotopic ratios were more depleted. This effect is more pronounced during periods with low precipitation, because of the evaporative enrichment of raindrops as a consequence of lower humidity. Conversely, high intensity and heavy rainfalls 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.

2.4.5 Temperature relationship

There is an apparent relationship between the local air temperature and the isotopic ratio of the precipitation. Dansgaard (1964) suggested on the basis of a limited database, an average relationship of 0.69‰/°C for ^{18}O and 5.6‰/°C for D. Similar values were also predicted by van der Straaten and Mook (1983), whose theoretical considerations were based on Raleigh condensation model with isobaric cooling of precipitating air masses. The studies on a larger data base by Rozanski et al. (1993) substantiated the predictions that, for reduced condensation temperature the slope of $\delta^{18}O$ - temperature, the relationship $d\delta^{18}O/dt$ increases.

2.4.6 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 will take place during this process. The adiabatic lapse rate, though it varies with altitude, is about 0.6°C/100m. From Dansgaard (1964) it is seen that, for δ_{18} the temperature dependence during adiabatic cooling is about 0.5‰/°C. This means that, there may be about 0.3‰ variation in δ_{18} per 100 meter variation in altitude. Clark and Fritz (1997) have compiled the results of several studies by different investigators that incorporated the altitude effect in precipitation. The altitude gradient (per 100 metres) varies between -0.1 ‰ and -0.5 ‰ for $\delta^{18}O$ and -2.5 ‰ and -4 ‰ for δD.

The adiabatic lapse rate of the cloud mass and hence the altitude effect in 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 and that evaporation is greater for raindrops that fall further (Coplen, 1993). However, the altitude effects are often not observed for rainfall in the lee side of the mountainous terrain and also in case of precipitation in the form snow fall.

2.4.7 Relationship between δ_p and 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 by Yurtsever and Gat (1981), who proposed an equation of the following type:

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

where T, P, L and A are the monthly mean temperature (°C), monthly precipitation (mm), latitude (degrees) and altitude (m a.s.l.) respectively. a_0, a_1, a_2, a_3 and a_4 are regression coefficients. However, using P, L and A did not significantly improve the statistical parameters of fitness. Monthly composite $\delta^{18}O$ value can be generated by a simple linear regression with T as the independent variable. However, they cautioned that the findings are true only on a global scale, and suggested that on regional scale either amount effect or evaporation effect may be equally important in determining the spatial variability.

2.4.8 $\delta^{18}O$ - δD relationships

The striking relationship between $\delta^{18}O$ and δD in freshwaters was first noted by Friedman (1953). δD of freshwaters was plotted as a function of $\delta^{18}O$ by Craig (1961), who suggested a best fit line for the data. The best fit line is called as Craig's line and is described by the following equation:

$$\delta D = 8\,\delta^{18}O + 10 \qquad (2.10)$$

Currently the equation of the Global Meteoric Water Line (GMWL) constructed on the basis of the long-term weighted mean hydrogen and oxygen isotopic ratios collected through IAEA/WMO database is (Rozanski et al., 1993):

$$\delta D = 8.2\,\delta^{18}O + 11.27 \qquad (2.11)$$

The above equation is identical to the earlier equations of the GMWL proposed by Dansgaard (1964) and Yurtsever and Gat (1981) 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 (2.11) is only apparent and as such cannot be used to derive the isotopic ratio of hydrogen, given the isotopic ratio of oxygen for any location. This is because the GMWL is essentially an average of several Local Meteoric Water Lines (LMWL). Dansgaard (1964) classified six different sets of stations (IAEA/WMO precipitation collection stations which were then operating) on the basis of $\delta^{18}O$ - δD relationship. He further enumerated the reasons for deviations in the $\delta^{18}O$ - δD relationship at different stations viz. a) initial vapour isotopic composition b) initial dew point temperature c) degree of cooling d) way of cooling and e) kinetic effects during fall of rain drops.

The LMWL having slopes lower or higher than that of the GMWL are not uncommon. Deviations have been reported for Israel, Jordan, Continental Europe and even in some tropical islands. The equation for Indian monsoon, on the basis of extensive groundwater samples collected in the northern India, proposed by Bhattacharya et al. (1985):

$$\delta D = 7.2\ \delta^{18}O + 5.1 \qquad (2.12)$$

The equation (2.12) is distinctly different from the equation proposed, on the basis of long-term annual weighted mean values of isotope ratios in the precipitation samples collected at New Delhi, by Datta et al (1991):

$$\delta D = 8.4\ \delta^{18}O + 11.4 \qquad (2.13)$$

However, the equation for the long term weighted mean isotopic ratios for monsoon months (June-September), proposed by Datta et al. (1991), for the same station is close to equation (2.12)

$$\delta D = 6.8\ \delta^{18}O + 0.7 \qquad (2.14)$$

The variation in the slope as well as intercept clearly brings out the fact that the relation between $\delta^{18}O$ and δD is more complex in case of Indian monsoon precipitation.

2.4.9 'd' - excess parameter

This all important parameter introduced by Dansgaard (1964), also called as d-index, literally means the surplus deuterium relative to the Craig's Line. i.e.

$$d = \delta D - 8\ \delta^{18}O \qquad (2.15)$$

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 a long as exchange is not a dominating factor,

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

2.5 Lake - Groundwater Interaction Studies

The lake generally acquires a specific isotopic labelling due to the process of evaporation and will be significantly different from that of the adjacent groundwater system that is recharged by local precipitation. The lake will be comparatively enriched in heavier isotopes. Since, the evaporation of the lake does not always take place under equilibrium conditions the lake is said to be 'doubly labelled' in that the d-index of the lake will also be different compared to the groundwater. Although, the evolution of isotopic composition of a lake depends up on several processes such as channelised and sheet form surface inflow, subsurface 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 (1965) Linear resistance Model:

a. Release of water vapour in isotopic equilibrium with liquid phase. Gat(1981) proposed the existence of a saturated sub-layer at the water-air interface, which is, compared to the liquid phase, depleted in the heavier isotope by ϵ^+.

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 evaporation model by Craig & Gordon (1965) can be written as:

$$\delta_E = \frac{(\alpha^* \delta_L - h\delta_a - \epsilon)}{(1-h+10^{-3}\Delta\epsilon)} \qquad (2.16)$$

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, α^*, ϵ and $\Delta\epsilon$ are as given in sections 2.3.2.1 and 2.3.2.2.

Suitable way to determine or by-pass δ_a have been attempted by several investigators. The different methods are discussed in the following section:

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. Dinçer (1968) suggested that samples could be collected at the point where relative humidity (RH) 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. He reasoned that the vapour produced by the lake has little influence on the isotopic composition of the net vapour flux from the lake. Krabbenhoft (1990) collected samples from lake's windward shore, by pumping 1.5 lpm of air through a liquid nitrogen trap. The method of collection has also been outlined by many investigators. A description of such samplers has been presented by Zimmermann (1972). Gat (1970) used a cold trap to collect atmospheric water vapour for δ_a determination. Zimmermann (1979) used an automatic water vapour sampler during the water balance study of Waidsee and Wiesensee lakes (Germany).

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 best suited for the purpose. In these cases, cost could be a constraining factor.

Index Lake Method:

The method introduced by Dinçer (1968), using a deep terminal lake, with constant isotopic composition, and evaporation equals inflow. The assumptions - region of uniform climatic characteristics, no significant aerial variation in isotopic composition of atmospheric water vapour, surface temperatures of the lakes are close to the atmospheric temperature, relative humidity has no large variation - were made by Dinçer in adopting the method. Since the lake is terminal type, that is the loss of water is only through evaporation, the following relationship holds:

$$\delta_E = \delta_I \tag{2.17}$$

For a terminal lake which has reached both hydrologic and isotopic steady state, equation (2.17) could be substituted in equation (2.16) and δ_a be determined, which in turn can then be used for other nearby lakes. This method requires the existence of an index lake which meets all the assumptions made by Dinçer, as outlined above, near the lake or lakes 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:

Welhan and Fritz (1977) proposed that for an isolated water body such as an evaporation pan, the following expression is valid:

$$(\delta - \delta_s)/(\delta_o - \delta_s) = f^m \tag{2.18}$$

and

$$m = (h - \epsilon)/(1 - h + \Delta\epsilon) \tag{2.19}$$

where, $f = V/V_0$ (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 (2.18) can be rewritten as,

$$\ln(\delta - \delta_s) - \ln(\delta_o - \delta_s) = m \cdot \ln f \tag{2.20}$$

so that the slope of the best fit curve of ln(δ - δ_s) vs. ln (f) will be m_{pan} and δ_s could be determined on trial basis.

If the mean lake-surface temperature and pan temperature do not vary much, by modifying Craig & Gordon's equation and assuming negligible error in applying m and δ_s of a pan to the lake, δ_E could be determined by using the following equation:

$$m_{pan} = [(\delta_E)_{lake} - \delta_{lake}] / [\delta_{lake} - (\delta_s)_{pan}] \qquad (2.21)$$

Stolf and co-workers (1979) have used a leaky evaporation pan theory somewhat close to this, to investigate the water balance of a reservoir in Brazil.

However, since the pan evaporation seldom proceeds under steady environment conditions to facilitate these computations, a modified version of this approach was proposed by Allison et al. (1979).

Multiple Pan Method:

Allison et al. (1979) attempted to determine the effect that the environment of a pan has on those parameters which have the most influence on the value of δ_E. They 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 than precipitation influenced breaks in δD - f ($f=V/V_o$) 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 the location of the pan is not an important criteria under normal conditions. The pan methods, are yet to be proved on their applicability to different environments. But these methods may be tried, if the relative humidity does not show a large variability.

δ_p equilibrium assumption method:

The relationship between isotopic composition of atmospheric water vapour and precipitation was reported by Zimmermann *et al.* (1967). Craig & Horibe (1967) reported that the isotopic composition of continental water vapour, on rainy days, is in equilibrium with that of precipitation. This implies that there could also be a period when they are not in equilibrium. Zuber (1983) assumed that δ_a is in equilibrium with the mean isotopic content of precipitation, so that

$$\delta_a = \alpha^* \delta_p - e^* \tag{2.22}$$

The study by Krabbenhoft *et al.* (1990) lend support to this assumption. The measured δ_a values compare well with those computed using equations (2.2) and (2.22) for measured RH and air temperature. Krabbenhoft et al. have opined that the deviations (about 2.5‰) of observed δ_a from calculated δ_a for the warmer months during their study period were possibly due to contributions from numerous lakes situated nearby the lake studied by them. Overall, given the constraints, it seems to be reasonable that if short-term δ_p is available, this method could be used to estimate δ_a.

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. Gat (1981) has presented the values of hydration fractionation factors of some electrolytes. A third salt effect is due to crystallisation of salts, which is minor one. 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 very last stages of evaporation. However, Gonfiantini (1986) noted that the salt effect on evaporating water bodies can be neglected in most natural cases.

2.5.2 Quantitative evaluation of the lake - aquifer interaction

The lake water - groundwater interaction can be quantitatively evaluated by focusing on the lake system by using the mass balance approach. One of the basic assumptions in the mass balance approach is that the lake is a well mixed one with uniform isotopic composition. However, lakes that do not confirm to this assumption do exist and they have been presented 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 is gaining substantial quantity of water from the adjacent aquifers is called as "discharge lakes" or "seepage lakes", and a lake that loses water to adjacent aquifer is called as "recharge lake" or "groundwater lake". Lakes that lose water only through evaporation is called as "terminal lakes". There are variations to this definitions if a lake is having no inflow or if the inflow compensates the evaporation. A lake that has substantial subsurface inflow and subsurface outflow is called as a "flow-through lake". Gonfiantini (1986) has presented a range of lake conditions and the equations which describe the evolution of the isotopic composition of the lakes, which can be used to compute the isotopic mass balance. Some probable lake types have been presented below:

Case I: Flow-through lakes

The evaluation of lake - aquifer interaction can be achieved, by simultaneously solving the material balance and isotope mass balance equations. The conventional water balance equation for the lake may be written as:

$$\Delta V = I_s + I_p + I_{ss} - O_s - O_e - O_{ss} \qquad (2.23)$$

where, ΔV is the change in volume, I_s surface inflow, I_p is direct precipitation on lake, I_{ss} is sub-surface inflow, O_s is surface outflow, O_e is evaporation from lake surface and O_{ss} is sub-surface outflow. Established conventional techniques can be used to measure or estimate all the above components within certain error limits (Winter, 1981) except the sub-surface ones. Although seepage meters have been used (Lee, 1977), 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

$$\frac{d(\delta_L V)}{dt} = \delta_s I_s + \delta_p I_p + \delta_g I_{ss} - \delta_L O_s - \delta_e O_e - \delta_L O_{ss} \qquad (2.24)$$

where t is the time period for which the balance is being computed, I and O with different subscripts are as given in equation (2.23), and δ-notations are the corresponding isotope ratios. The equation has been simplified by assigning the lake δ_L to the outflow components ($\delta_{o\text{-}s}$, $\delta_{o\text{-}ss}$) and δ_g to subsurface inflow ($\delta_{i\text{-}ss}$). In this equation except δ_E, all other components are directly measurable.

Zuber (1983) presented the following equation to estimate the outflow to evaporation ratio, which can then be used with conventional data to estimate the subsurface outflow components:

$$x = \frac{O}{O_E} = \frac{\left(\frac{I_S}{O_E}\right)(\delta_G - \delta_S) + \left(\frac{I_P}{O_E}\right)(\delta_G - \delta_P) + \delta_E - \delta_G}{(\delta_G - \delta_I)} \qquad (2.25)$$

He suggested that the above equation (2.25) may further be simplified by assuming δ_G is equal to δ_P, in the absence of groundwater data. Zuber (1983) calculated the subsurface outflow from

Lake Chala, Tanzania using oxygen isotope ratios. He reported a value of 3.9 * 10^6 m^3/yr, which was in very good agreement with the value of 4.14 * 10^6 m^3/yr reported by Payne (1983), who used injected artificial tritium for whole body tracing.

In the study of Lake Bagry, Poland also, Zuber (1983) used equation (2.25) to estimate the subsurface inflow and outflow of the lake. 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, Zuber used the mean value by integrating the area below δ_L - time curve. The residence time (V/O) calculated by using the results of isotope method compared well with that obtained through electrical analogy approach, which confirmed the validity of the isotope approach, as adopted by Zuber.

Krabbenhoft et al. (1990) estimated the groundwater exchange with Lake Sparkling, Wisconsin, USA using the stable isotope mass balance approach using oxygen isotopes. Since the surface inflow to the lake was insignificant and the lake was in a isotopically steady state condition they derived an equation from equations (2.23) and (2.24) that satisfies the above observations, such that:

$$I_{ss} = \frac{I_P(\delta_l - \delta_p) + O_E(\delta_E - \delta_l)}{\delta_g - \delta_l} \tag{2.26}$$

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

The Sparkling lake study was later extended by Krabbenhoft et al. (1994), considering the lake to be a Index lake, to study the groundwater components of other nearby lakes with negligible surface components. They studied the groundwater exchange of Lake Crystal, Lake Pallette, Lake Big Musky. Assuming isotopic steady state of the lake, they used the groundwater inflow information of Sparkling lake to evaluate the subsurface components of the three lakes. The success of this approach depends very much on the accuracy of the estimation of the water balance components of the Index lake.

LaBaugh et al. (1997) estimated the groundwater component in the water balance of Williams lake, Minnesota, USA using a multi-tracer approach and flow net approach. The investigators found a large difference between the results obtained through isotope approach and flow net approach. They reasoned that the discrepancy arose mainly due to the errors attached to 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 (seepage) and the range of uncertainty varied with the selection procedure of the groundwater isotopic index. There was a difference of ~2‰ between the isotopic composition groundwater samples from the seepage zones 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 their study area. They concluded that the oxygen isotope was a better tracer in accounting the exchange of waters of lake Williams with groundwater than the four chemical tracers viz. chloride, sodium, magnesium or dissolved organic carbon that were tried. Further, they suggested that the combined use of hydrogeological and chemical approaches will improve the estimation of the lake - aquifer exchanges.

Case II: Desiccating terminal lakes with no inflow:

For a terminal lake with no inflow, the variation of the isotopic composition could be written as (Gonfiantini, 1986):

$$\frac{d\delta_l}{d\ln f} = \frac{h(\delta_l - \delta_a) - (\delta_l + 1)\left(\Delta\epsilon + \frac{\epsilon^+}{\alpha^+}\right)}{1 - h + \Delta\epsilon} \qquad (2.27)$$

Where f is the fraction of remaining water. Assuming environmental conditions of evaporation as constant i.e., h, δa, $\Delta\epsilon$ and ϵ^+ as constant, equation (2.27) can be integrated to get

$$\delta_L = \left(\delta_{L_o} - \frac{A}{B}\right) f^{B} + \frac{A}{B} \qquad (2.28)$$

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

$$A = \frac{h\delta_a + \Delta\epsilon + \frac{\epsilon^*}{\alpha^*}}{1 - h + \Delta\epsilon} \qquad (2.29)$$

$$B = \frac{h - \Delta\epsilon - \frac{\epsilon^*}{\alpha^*}}{1 - h + \Delta\epsilon} \qquad (2.30)$$

From equation (2.28) it is seen that as f tends to zero (during final stages of evaporation) the term (A/B) is then the final isotopic composition of the lake and it is independent of initial lake isotopic composition.

Case III: Lake with slow and unidirectional change in volume:

For lakes / 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 also the inflow, outflow and evaporation rates to be constant (Gonfiantini, 1986):

$$\delta_l = \left(\delta_{l_o} - \frac{\delta_l + AX}{1 + BX}\right) f^{\frac{-(1 + BX)}{(1 - X - Y)}} + \frac{\delta_l + AX}{1 + BX} \qquad (2.31)$$

where, X = E/I and Y = Q/I.

Case IV: Lakes of leaky evaporation pan type:

This type of the lake is similar to the one given under Case III, but with no inflow. i.e., I = 0.

$$\delta_l = \left(\delta_{l_o} - \frac{A}{B}\right) f^{Z} + \frac{A}{B} \qquad (2.32)$$

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, Stolf et al. (1979) used this approach. Since there was negligible surface inflow and no surface outflow to the reservoir, the investigators used the leaking evaporation an analogy to estimate the evaporation and subsurface outflow from the reservoir. Equation (2.32) was used with the oxygen isotope data of the pan form which the values of parameters A and B were evaluated, which were then reused in the equation (2.32) to evaluate the parameter Z of the reservoir. The investigators estimated that ~67% of the total losses from reservoir was accounted by evaporation and ~33% by subsurface outflow. An independent tracer, viz. chloride ion was also used for the study which indicated that subsurface outflow accounted for ~25% of total losses. However, the investigators reported that the results of two independent tracers are in reasonable agreement, if the error associated to the computations are taken into consideration.

Case V: Lake 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 to evaporation and V/I = T, where T is the mean residence time of the lake we get:

$$\frac{\delta_l}{\delta_t} = -\frac{1}{T}[(1 + BX)\ \delta_l - \delta_i - AX] \qquad (2.33)$$

Integrating equation (2.33) and considering evaporation conditions and the isotopic composition of inflow waters to be constant, we get:

$$\delta_L = \left(\frac{\delta_i + AX}{1 + BX} \right) + \left(\delta_{L_o} - \frac{\delta_i + AX}{1 + BX} \right) e^{-(1 + BX)\frac{t}{T}} \qquad (2.34)$$

In the above equation when t tends to infinity, δ_L tends to δs, a steady state value is then achieved, which is defined by:

$$\delta_s = \frac{\delta_i + AX}{1 + BX} \qquad (2.35)$$

The groundwater components in the water balance of two young artificial lakes in Germany viz. Lake Wiesensee and Lake Waidsee were investigated by Zimmermann (1979). The lake had no surface inflow or outflow. Zimmermann considered the lakes to be of constant volume, although there was a 6% decrease in the volume. During the investigation period from 1970 - 1974, the lakes showed a continuous enrichment in the heavy isotope (δD) due to evaporation, on which the seasonal variations in the precipitation were superimposed. The investigator correlated $\ln(\delta_L{}^S - \delta_L)$ with time (t), and selected the value of $\delta_L{}^S$ by trial basis using equation (2.34). The values of parameters A and B were computed using the field humidity and δ_a values. In the study, the results attached to the estimated evaporation and subsurface inflow and outflow were of the order of 30%. The investigator, however, concluded that the stable isotope method is suitable for computing subsurface components as there is no other method to evaluate these components with comparable accuracy.

Zuber (1983) using equation (2.25) reinterpreted the data on Lake Waidsee presented by Zimmermann. Based on the discrepancy in the value obtained for X, Zuber questioned the results of Zimmermann (1979). Further, Zuber showed that the assumptions involved in the calculations influences the results of the parameter X.

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 could be determined from (Gonfiantini, 1986):

$$\delta_S^T = \alpha'\delta_i \, (1 - h + \Delta\epsilon) + \alpha' h \delta_a + \alpha'\Delta\epsilon + \epsilon' \qquad (2.36)$$

δ^T_S is then the maximum enrichment in heavy isotope which can be reached by a constant volume lake. Correction terms have to be included in equation (2.36) as in certain terminal lakes the salt effects may dominate.

Dincer (1968) used the presence of Lake Burdur, in Turkey which is of terminal type, as an index lake for molecular exchange in the region where two fresh water lakes viz. Lake Egridir and Lake Beysehir were situated. He used the following equation to estimate the water balance of the fresh water lakes:

$$\frac{d(\delta_l V)}{dt} = I \, (\delta_i - \delta_l) - E \left(\frac{\delta_l - \delta_l^T}{\alpha \, (1 - h)} + \delta_i^T - \delta_l \right) \qquad (2.37)$$

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 difficult, however with a dense network of sampling and using a properly weighted mean value to solve the mass balance equations with a reasonable degree of accuracy may be attempted. Fontes et al. (1970) utilised the systematic variation in the isotopic composition in a shallow lake, disregarding mixing and considered each sample as representative of different stages of evaporation of the lake. While studying a large swamp area, Dincer (1979) combined isotopic and water balance equations with the salt balance equations. They suggested that the method allowed the calculation of the ratio of evaporation to evapotranspiration.

Gonfiantini et al. (1962) investigated the water balance of lake Bracciano, Italy using isotope approach. 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. Zimmermann and Ehhalt (1970) used hydrogen isotope ratio to study the subsurface components of Lake Neuseidl, Austria, which is a shallow lake with poor mixing. The investigators neglected the kinetic enrichment effect in their computations. They concluded that large errors associated with δ_E resulted the determination of evaporation rate not better than 50% accuracy.

The case of the seasonal stratification is not uncommon in temperate lakes, in 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 (by evaporation) compared to the bottom waters, called hypolimnion. In the water balance study of Lake Tiberias, which is a stratified lake, Gat (1970) used 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. Gat (1995) suggested that the effect due to exchange with hypolimnion is usually not significant, if exchange is the only process by which the isotopic composition of the hypolimnion changes. However, the effect of hypolimnion entrainment will generally be significant.

2.5.3 Qualitative studies on lake - aquifer interaction

Isotope methods are also useful in accepting or rejecting hypothesis of lake water interaction with local groundwater or springs issuing at greater distances away from the lakes.

Stichler and Moser (1979) used the stable isotope techniques to find the direction of subsurface outflow of a lake in Germany. Payne (1970) disproved the belief that certain springs in the vicinity of lake Chala, Kenya are fed by the lake. Payne showed that the lake's contribution to the springs is negligible, on the basis of stable isotope data. Recently, Abidin et al. (1996) used oxygen isotope ratios to establish the connection between the Mt. Galunggung crater lake in Indonesia and a nearby spring.

3.0 CHARACTERISTICS OF LAKE NAINITAL BASIN

3.1 Geology

The area stretching between the flat plains of Ganges basin and the mature old terrain of the lesser Himalayas, standing out as a physiographic eminence to more than 2500m in height, is the Nainital Massif. The Nainital Massif may be defined as the mountain front of Himalayas in Kumaun region and it (the Nainital Massif) forms the southeastern extremity of the Krol belt that extends for about 300 km from Solan in Himachal Pradesh.

Valdiya (1988) describes the Nainital Massif as the resultant of the phenomenon that makes the very old rocks of Himalayas (2000 - 1800 to 600 - 500 m.a.) advance over the very young Siwalik (20 - 2 m.a.) along the 2400 km long regional thrust called the Main Boundary Thrust. The phenomenon has also resulted in severe compression of the Massif into highly overturned folds split by numerous fractures and faults. Differential movements of fractured blocks have given rise to horsts and grabens in the buckled-up terrain which are aptly described by Valdiya (1988) as block-faulted mountains.

3.1.1 Lithostratigraphy

The great succession of Nainital Massif is divisible into lower argillo-arenaceous Jaunsar Group and upper argillo-calcareous Mussoorie Group (Valdiya, 1980b). The stratigraphic succession of the Mussoorie Group that characterises the Lake basin has been presented by Valdiya (1988):

Table 3.1. Lithostratigraphic succession of the Mussoorie Group in Nainital Hills of the Krol Belt, Kumaun Lesser Himalayas (Valdiya, 1988)

Mussoorie Group	Tal Formation	NarainNagar Member	= Middle Tal	
		Giwalikhet Member	= Lower Tal	
	Krol Formation	Sherwood Member	= Krol E	
		Bist College Member	= local facies	
		Pashandevi Member	= Krol D	
		Barapatthar Member	= Krol C	
		Hanumangarhi Member	= Krol B	
		Manora Member	= Krol A	
	Blaini Formation	Kailakhan Slate	= Infra Krol	
		Pangot Member	= Blaini	

Blaini Formation: The Blaini Formation is divisible into two distinct members - the lower conglomeritic calc-arenaceous unit and upper argillaceous horizon. The underlying Lariakantha quartzite of Jaunsar Group is succeeded without structural discordance by Pangot Member- a succession of diamictite intimately associated with purple slate, calcareous sandstone and siliceous-dolomitic limestone of deep pink colour. The horizon is best developed in the Kilberry-Pangot-Gajkhet belt NNE of Nainital town. The purple slates associated with the Pangot diamictite give way to Kailakhan Slate - the ash-grey or black shale. Occurrence of pyritic and phosphatic nodules has been noted in areas such as on the east-facing slope of Hanumangarhi temple. The Kailakhan Slates are best seen on the Kailakhan-Alukhet spur, southeast of Nainital town.

Krol Formation: The carbonate rocks in and around the Lake Nainital have been identified as the Krol by several geologists (Middlemiss, 1890; Heim and Gansser, 1939; Auden, 1942; Awasthi, 1970; Hukku et al., 1974; Pal and Merh, 1974; Pande, 1974; Fuchs and Sinha, 1974; Valdiya, 1979, 1980b, 1981 and 1988; Sharma, 1981). The Krol Formation is divisible into six members as clearly seen in the Tiffin-Top Baldiakhan section across Ayarpatta ridge. Auden(1934) and Bhargava (1976) have noted that with the exception of a lone local horizon

(Bist College Member) the lithological succession and its subdivision in Nainital Hills is the same as in the Mussoorie Hills.

The Manora Member is characterised by light yellowish, green-grey and light brown calcareous slates alternate with marlite and argillaceous limestone. The horizon is well developed on the Manora ridge and can be seen in the Cantonment area on the Nainital town- Bhowali area. The Hanumangarhi Member is characterised by red and purple ferruginous shales, locally interbedded with subordinate but conspicuous algal dolomites forming biohermal lenses. The horizon is well developed in the Hanumangarhi temple area. Pockets of gypsum are very characteristic as seen in Nihal Valley. Barapathar Member is characterised by rhythmic alternation of thin laminae of black shales with marlite or argillaceous - carbonaceous limestone. It grades into richly carbonaceous shales. The horizon is best developed at Barapathar on the Kaladungi road, near Sariyatal and on the western slope of the Hanumangarhi temple.

The Pashandevi Member that is the lowest unit of the Upper Krol is characterised by massive dark-grey and blue dolomitic limestone. A large part of the northern slope of the Ayarpatta including the Pashandevi spur on the lake shore is made up of this horizon. The Bist college Member is a local lenticular facies of the Upper Krol semicircling the Ayarpatta area. The Member is characterised by greywacke and siltstone of purple, maroon, brown, grey and fawn colours with subordinate shales of like colours. The sandstone is locally ferruginous, sericitic and commonly fine-grained. It is traceable from the D.S.B. Campus of Kumaun University, through the District Court and Craigland to west of Rajbhawan. The Sherwood Member is characterised by massive dolomite which is locally cherty and phosphatic. the P_2O_5 content is around 17-30% (Choudhary and Banerjee, 1966). The horizon is well developed in the summit of the Ayarpatta ridge. The Sherwood Member transitionally grades into the Giwalikhet Member of Tal Formation.

Tal Formation: The succession of carbonaceous shale with subordinate dolomitic limestone characterised by nodules, laminae and stringers of phosphatic material, followed upward by purple-green shales intercalated with muddy fine-grained sandstone and siltstone is identified as Tal formation (Valdiya, 1980a) that is divisible into two members. The lower

Giwalikhet Member, best developed in the Golf course of Rajbhawan, is comparable with the Chert-phosphorite and Argillaceous Members, and the upper NarainNagar Member with the arenaceous Member of the Mussoorie hills (Shanker, 1971; Bhargava, 1976, 1979; Kumar et al., 1983).

Dolerite: The dolerite that occur as dykes has been described as a intrusive trap dyke by Middlemiss (1890). The dyke occurs in the west, southwest and south of Ayarpatta peak (Pande, 1974). The dyke is also seen in the Alma lodge area between Marshall cottage and Snow view. Middlemiss (1890) did not rule out the possibility of subterranean connection between the two dykes across the lake. The bathymetric map of the Lake Nainital (Hashimi et al., 1993) does provide some clue regarding the possibility of the dyke in the Southeastern side of the Lake Nainital extending upto the Nainital Fault.

3.1.2 Structure

A regional fault - called as Lake fault or Naini fault - extending from Jeolikot through Nainital lake to Kunjakharak cuts the deformed synclinal massif into two dissimilar parts (Figure 3.1). The narrower southern part is made predominantly of younger group of sedimentary rocks (Blaini, Krol and Tal formations) and is characterised by tight folds that verge southwards. The larger northern part is made predominantly of older Jaunsar group of rocks and is characterised by folds that are overturned towards north.

The structural geology of the lake basin and surroundings has been studied in detail by several geologists. (Sarkar et al. 1967; Pande, 1974; Pal and Merh, 1974; Sharma, 1981). The studies show that there are three phases of folding. The first phase of deformation resulted in larger folds trending NW-SE, that is exemplified by southwardly overturned asymmetric Ayarpatta - Gairkhet syncline and also by asymmetric and faulted Deopatta anticline. The second phase resulted in E-W to ENE-WSW oriented folds that is exemplified by Naina peak syncline. The third phase resulted in NE/NNE - SW/SSW trending gentler folds.

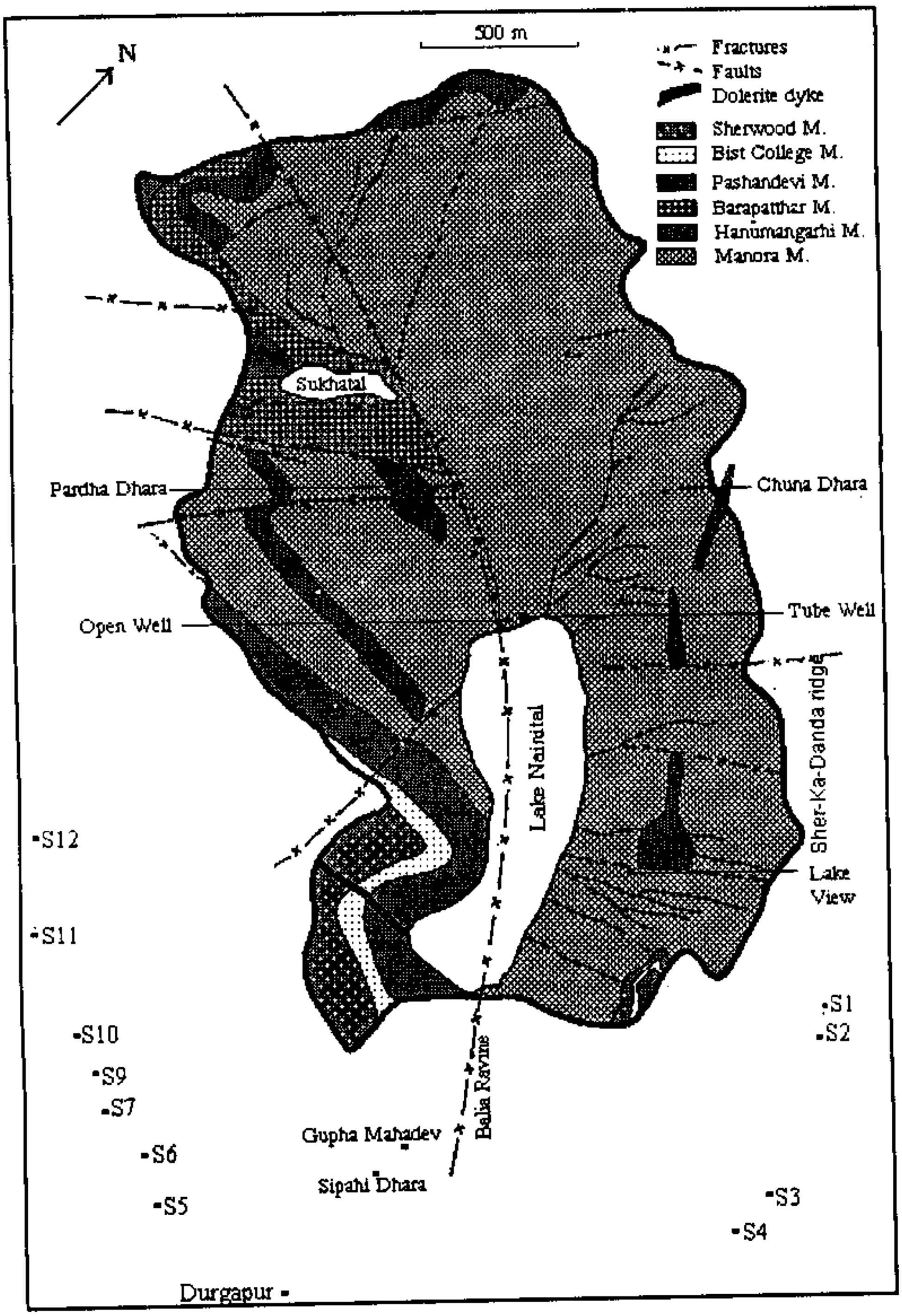


Figure 3.1 Geological map of Nainital basin showing locations of springs and wells

The steeply reversed regional Nainital fault, referred above, was first described by Middlemiss (1890). The fault is traceable from the saddle between Naina and Deopatta peaks through the Nainadevi temple along the Balia ravine. The fault has dextrally offset the Main Boundary Thrust near Beluwakhan, west of Jeolikot (Valdiya et al., 1984) and extends for a kilometre into the Siwaliks where it is delineated by a line of springs and landslides of Basgaon. Pande (1974) estimated the uplift of Naini peak - Sher-ka-Danda block in relation to the Ayarpatta block to be of the order of 250 m. According to Valdiya (1988) the Nainital fault, that developed close to or along the crestal plane of an overturned anticline, has considerably truncated the limbs of the anticline and brought the Sher-ka-Danda and Ayarpatta synclines in juxtaposition. The northwesterly extension of the intervening anticline is represented by the Deopatta.

Parallel to the Nainital fault, a sympathetic Ayarpatta fault (Middlemiss, 1890) has considerably deformed the overturned limb of the asymmetrical syncline lying south to it. The bedding plane fault developed between the Pashandevi dolomites and the Barapathar rhythmites is traceable between St. Andrews and Emily cottage. According to Valdiya (1988) the whole belt southeast of Nainital Fault is cut by a large number of N-S and NE/ENE-SW/WSW trending tear faults. The left-lateral Khurpatal fault described by Pal and Merh (1974) is probably linked with the Sleepy Hallow Fault (Valdiya, 1988) defined by the vertical scarps overlooking the Muslim cemetery and by the straight steep slope that demarcates Sleepy Hallow. The Sleepy Hallow Fault and another fault, that is traceable from NarainNagar through the Nainital Polytechnic, join the Nainital Fault. These two minor faults have offset and complexly drag folded the Middle Krol Unit in the Priori-Aroma area and given rise to a lakelet behind Snowdon.

3.1.3 Geochemistry of the rock formations of the Lake basin

From the available literature and the foregone discussions it is seen that the chief rock types of the lake basin are:

1. Argillaceous - Shales / Slates
2. Dolomites

3. Dolomitic Sandstones
4. Purple sandstones and quartzites
5. Dioritic traps

No detailed information is available on the geochemistry of the rocks occurring in the Nainital basin except for some data on the composition of a few samples collected from the lake basin by Holland (1896) on the basis of specific gravity of the samples. The average specific gravity of the blue-grey dolomite of Nainital is 2.82 that corresponds to 64.3% of $CaCO_3$ and 35.7% of $MgCO_3$. However the foreign matter are often considerable in quantity as shown by the chemical analysis of a specimen from Ayarpatta, which showed $CaCO_3$ - 50.13%, $MgCO_3$ - 40.89%, Alumina, etc. - 1.63% and Insoluble residues - 7.71%. Holland (1896) also described the presence of sandy dolomite & dolomitic sandstone near the East Laggan, the composition of which are as follows:

Table 3.2. Chemical composition of dolomitic rocks of Lake Nainital basin

	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%	-

Bagati (1974) reported on the geochemistry of the carbonate rocks of the Krol formation of Simla Himalayas that the Si content increases with increase in the insoluble residue. He further noted that the Ca/Mg ratio vary from 0.99 to 14.99 and that Ca/Mg ratio is higher for Limestones (14.46 - 14.99) and is lower for Dolomites (0.99 - 2.46).

3.2 Precipitation

The observation of rainfall in the Nainital basin started in the year 1895 and records of total annual rainfall are available continuously from 1895 onwards in the published reports of United Province Public Works Department, that was later reorganised as Uttar Pradesh Public Works Department (UPPWD). Presently the Manual Raingauge of the UPPWD is located in the UPPWD campus at SnowView. Further at Manora, located about 2 kilometres south of the Lake basin, Uttar Pradesh State Observatory (UPSO) has set up a meteorological station where the rainfall records are available since 1965. In addition to the above raingauges, three additional Self Recording Rain Gauges (Syphon Type) were installed with in the Lake basin by National Institute of Hydrology (NIH) to account for the spatial variations in the precipitation. The raingauge sites are ATI campus near Sukhatal, Rattan Cottage in the Sher-ka-Danda side and Alma Cottage. The data collected from UPPWD, UPSO and that collected from the NIH raingauges have been used to understand the precipitation characteristics in the study area. The locations of the raingauge stations are given in Figure 3.2.

3.2.1 Long-term average precipitation rates

The average annual precipitation at a site is calculated by totalling the precipitation catch for a number of complete years and dividing the sum by the number of years used. The 30-year normal precipitation is computed as the average annual precipitation for the specific 30-year period. Analysis of the UPSO rainfall record indicates that the average annual rainfall for the 30 year period (1966-95) is 2125 mm. The annual rainfall during the 30 year period ranges from a minimum of 1367 mm (1991) to a maximum of 3117 mm (1985). However, analysis of one hundred one year annual rainfall data available for Lake Nainital basin (UPPWD) indicates that the mean annual rainfall in the basin is 2488 mm, with a minimum of 1367 mm in 1991 and a maximum of 3910 mm in 1910.

Sample autocorrelation coefficients, that measure the correlation between observations at different distances apart, are an important guide to the properties of a time series. These coefficients often provide insight into the probability model that generated the data. For N

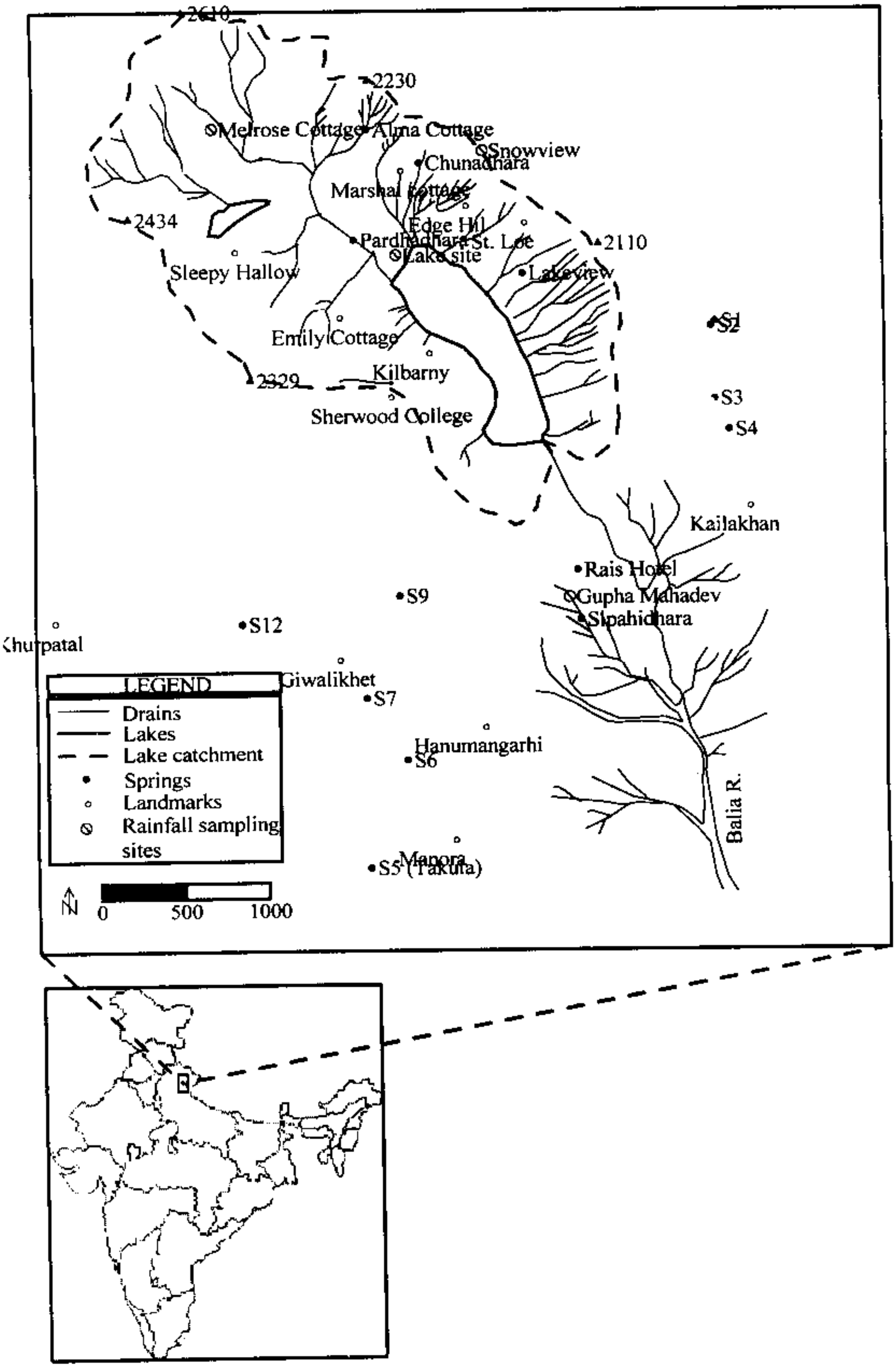


Figure 3.2 Location map of the study area showing Nainital lake and sampling sites

observations $x_1, x_2, x_3, \ldots\ldots x_N$ on a discrete time series one can perform (N-1) pairs of observations viz., (x_1,x_2), (x_2,x_3), $(x_3,x_4)\ldots\ldots(x_{N-1},x_N)$. Considering the first observation in each pair as one variable and the second one as a second variable, the correlation coefficients between x_t and x_{t+1} is given by

$$r_k = \frac{\sum_{t=1}^{N-1} (x_t - \overline{x_{(1)}})(x_{t+1} - \overline{x_{(2)}})}{\left(\sum_{t=1}^{N-1} (x_t - \overline{x_{(1)}})^2 \sum_{t=1}^{N-1} (x_{t+1} - \overline{x_{(2)}})^2 \right)^{1/2}} \tag{3.1}$$

where $x_{(1)}$ and $x_{(2)}$ are the mean of the first (N-1) and last (N-1) observations, respectively. Assuming that $x_{(1)} \approx x_{(2)} \approx x$, and that $N/(N-1)$ is close to one for large values of N, the above equation may be rewritten to find the correlation between observations at a distance k apart (Keshari and Kumar, 1997)

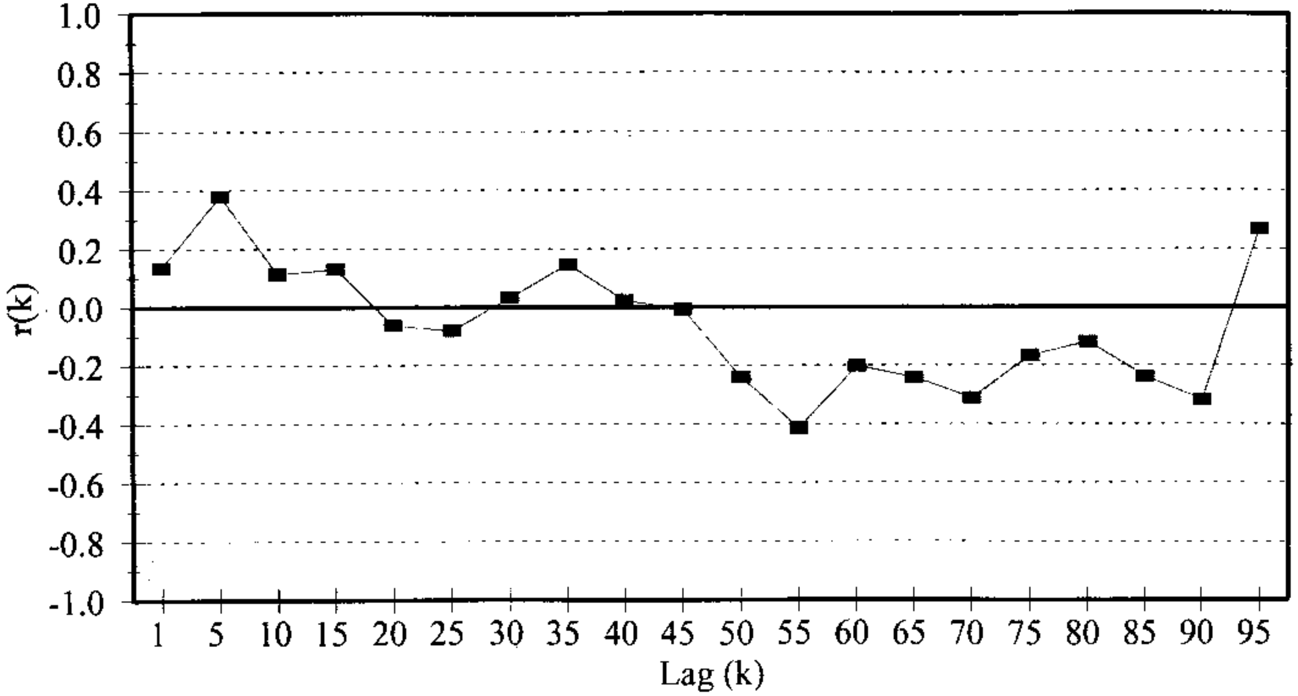
$$r_k = \frac{\sum_{t=1}^{N-k} (x_t - \overline{x})(x_{t+k} - \overline{x})}{\sum_{t=1}^{N} (x_t - \overline{x})^2} \tag{3.2}$$

where r_k is called the autocorrelation coefficient at lag k. One of the useful aids in the interpretation of the autocorrelation coefficients is the plot between the r_k and lag k, and the plot is called a correlogram. Time series analysis of the Nainital rainfall (Figure 3.3) shows that no trend, periodical or otherwise, is discernible in the annual rainfall series.

3.2.2 Seasonal variability

The seasonal variability of precipitation is an important aspect of hydrology because, it largely determines the seasonality of other hydrologic quantities such as stream flow and groundwater recharge. The seasonal patterns of precipitation process such as relative heating of the continents and the migration of large scale circulation features largely control the seasonality of precipitation on a global scale. Monsoon regions in particular have pronounced seasonal variation of precipitation. Most parts of the Indian sub-continent is characterised by the monsoon

Fig. 3.3 Correlogram of annual rainfall in Lake Nainital catchment



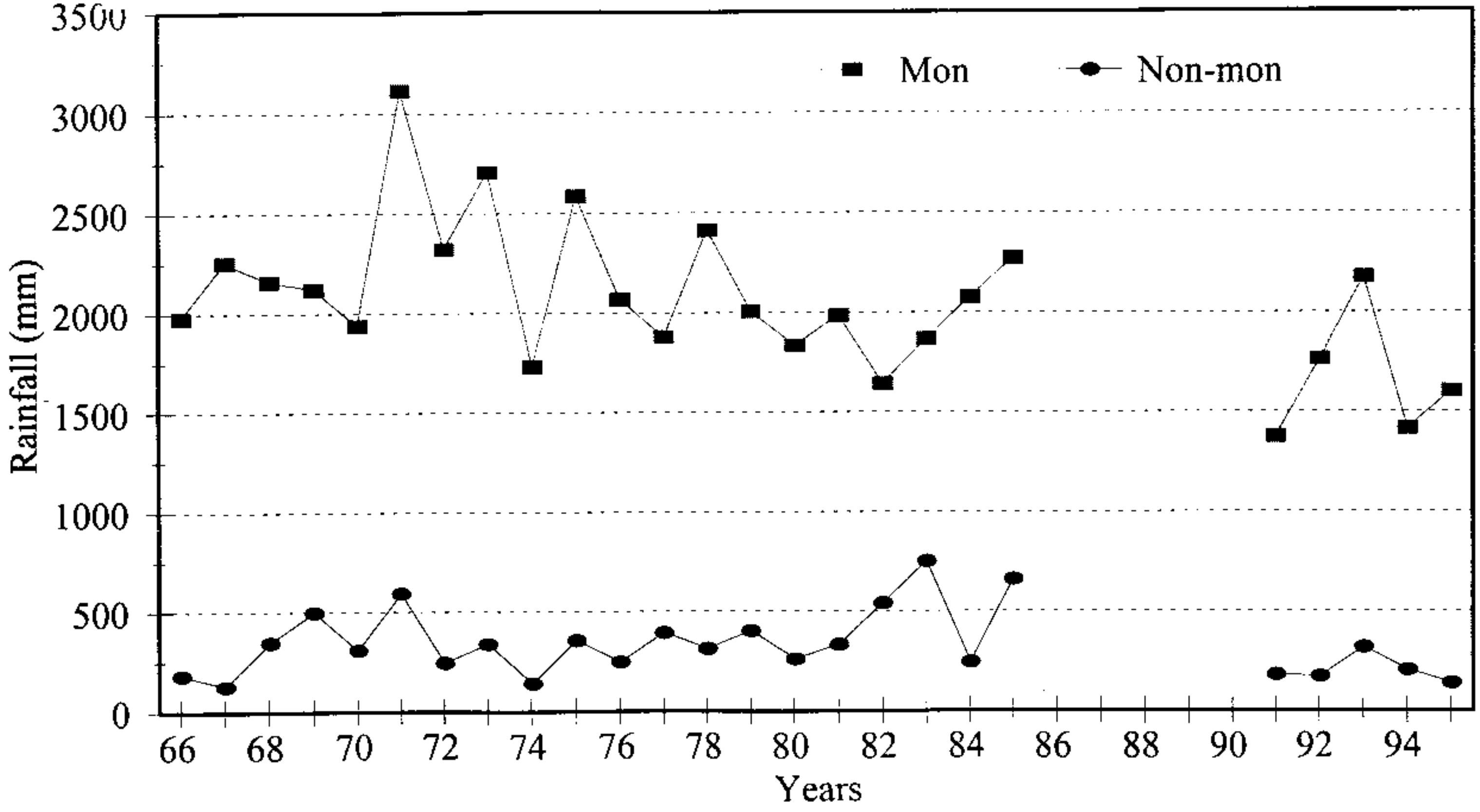
precipitation, and that the start and end dates of the monsoon season vary from place to place. The monsoon period (rainy season) in the Kumaun Himalayan region generally starts from 3rd week of June and extends up to September. During monsoon season the precipitation over the study area is mainly due to moist air-currents from Bay of Bengal. In general in the Himalayan ranges, the rainfall amount decreases westward because of increasing distance from the source of moisture. The precipitation during the monsoon season is received as moderate to heavy rainfall. In contrast to this the precipitation in the study area during winter season (January to March) is generally received as light to moderate rains with occasional snowfall, caused by extratropical weather systems of mid-latitude regions (originating over Caspian sea). These winter weather systems are known as Western - Disturbances.

From the 30 year average, it is seen that the monsoon rainfall contributes around 86% and non-monsoon rainfall contributes around 14%. The average monsoon and non-monsoon rainfall for the 30 year period is shown in Figure 3.4. This shows the monsoon rainfall plays an important role in the hydrology of the Nainital area.

3.2.3 Orographic effects on precipitation

It is a well-known fact that the interaction of weather systems with topography in mountainous basins results in non-uniform precipitation. Literature is abound with reports on the variation of rainfall amount and intensity with increase in altitude. Singh et al. (1995) on the orographic effect of precipitation in different ranges of Western Himalayas, observed that there is remarkable difference in the variation of rainfall with altitude among the three ranges viz., Outer, Middle and Greater Himalayas. On the rainfall distribution in the Outer Himalayas Singh et al. (1995) observed that, during monsoon season the rainfall in the windward side increases upto an altitude of 600m and then decreases, while during other seasons the amount of rainfall increases linearly with altitude. They attributed this differential behaviour to the differences in the weather conditions particularly moisture content in air, as during monsoon the moisture content in atmosphere and clouds are in saturated conditions and clouds become mature at relatively lower elevation.

Fig. 3.4 Variation of monsoon and non-monsoon rainfall at Manora (Nainital)



In the present study monthly rainfall data of three stations viz. Manora, ATI and SnowView have been used to know the orographic effect in the Nainital precipitation. It is seen from Figure 3.5 that the rainfall during July and August show a decreasing trend with altitude while the rainfall during September shows an increasing trend with altitude. This observation is substantiated by the variation in the 30-year mean monthly relative humidity (Figure 3.6) observed at Manora. The mean RH is higher during the months of July and August, and relatively lower during September.

3.2.4 Estimation of areal precipitation

Precipitation over a region, such as a drainage basin, rather than a point is of great importance for most hydrological purposes. several approaches have been derived for estimating regional precipitation from point values (for a specific time period). Theissen polygon method is a widely used one. The method involves division of the region into a specific number of sub-regions that are centred on each of the raingauges. The Theissen weights (ratio of each sub-region area to total area) are then assigned to each sub-region. This method yields better results when the gauge density is sufficiently high and that the gauges are distributed over the region.

The daily rainfall data collected through the three NIH raingauges installed within the Lake basin viz. ATI campus, Alma cottage and at Rattan cottage, along with that collected through the PWD raingauge at SnowView have been used in the Theissen polygon method, and the Theissen weights have been determined by measuring the area of the four sub-regions using a digital planimeter. The mean daily rainfall in the basin has been computed using the rainfall data collected from the above raingauges and the respective Theissen weights.

Figure 3.5 Orographic effect in Rainfall - Lake Nainital basin (1995)

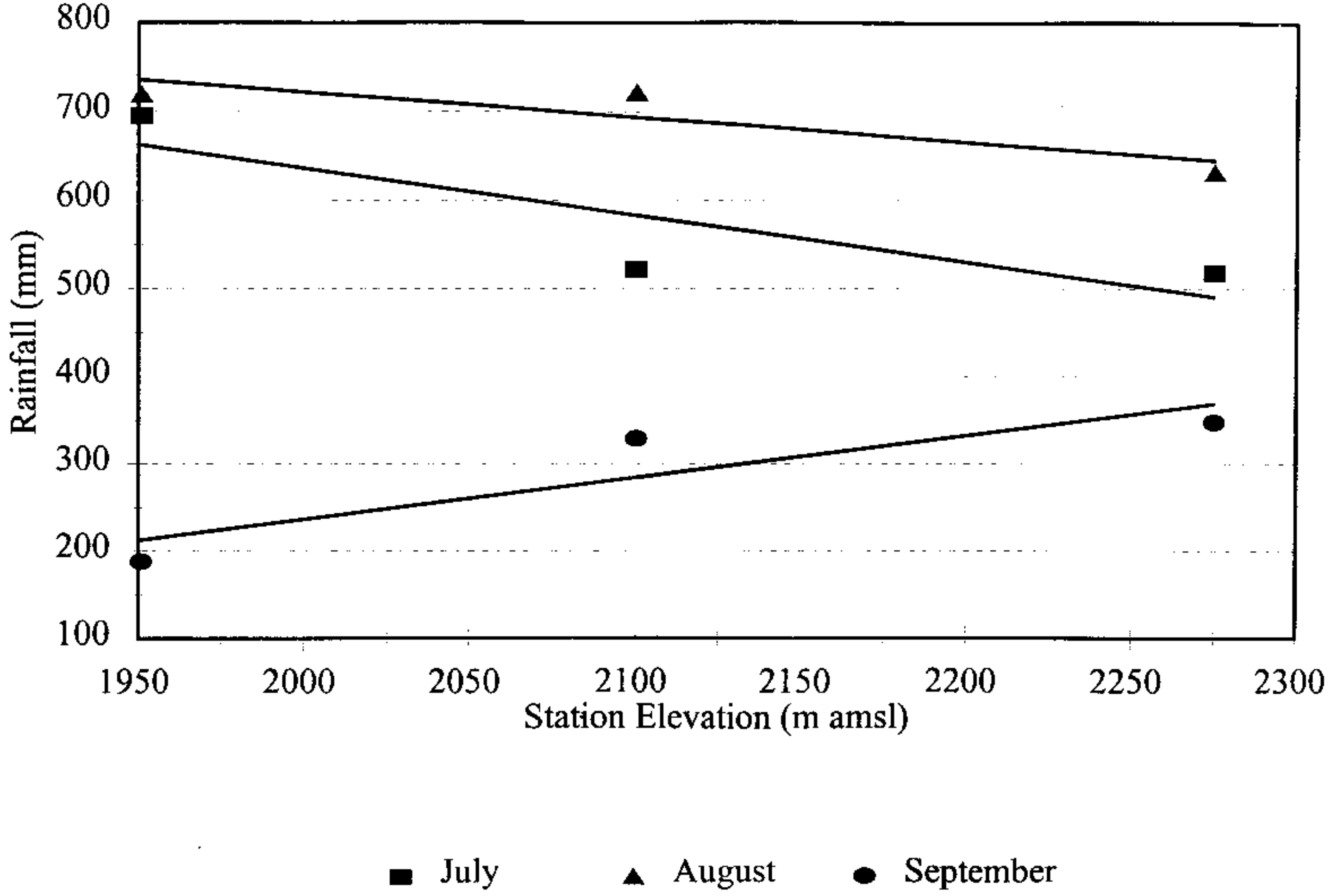


Fig. 3.6 Variation of 30 year average monthly mean relative humidity at Manora (Nainital)

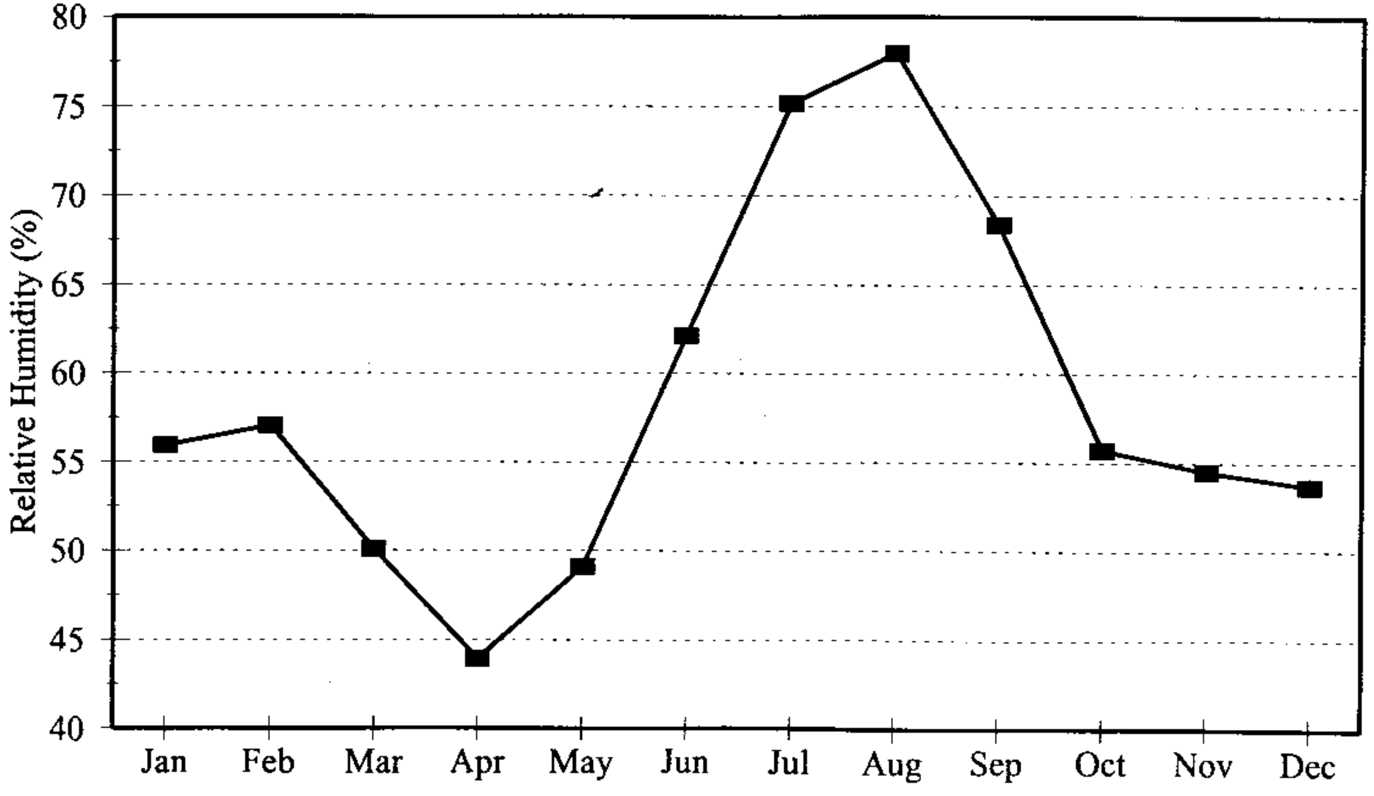


Table 3.3. Theissen Weights of different raingauge stations in Lake Nainital Basin.

Sl. No.	Rain Gauge station	Theissen Weights
1.	Alma Cottage	0.16
2.	ATI Campus	0.45
3.	PWD Site	0.14
4.	Rattan Cottage	0.25

3.2.5 Precipitation quality

Every raindrop and snowflake is formed around a nucleus, the most common of these are sea salt, wind blown dust, smoke particles and various pollutants. The relative abundances and solubilities of these materials are the first determinants of the quality of precipitation. Subsequently, the chemical composition of droplets change as they dissolve gases from the surrounding air at rates proportional to solubility and concentration of each gas. Berner and Berner (1987) have presented the typical values of the major ions in the continental rains. Under normal conditions rainfall is a very dilute solution dominated by Ca^{++} (~4 mg/L), Mg^{++} (~0.5 mg/L), Na^{+} (~1 mg/L), K^{+} (~0.5 mg/L), SO_4^{2-} (~3 mg/L), Cl^{-} (~2 mg/L), and NO_3^{-} (~0.5 mg/L) with a total concentration typically between 5 - 20 mg/L.

Since no actual observed data on the quality of the precipitation in Nainital basin is available, the same could be interpolated using the available data for the sites Lucknow and Srinagar (Jammu and Kashmir) that are the only stations closer to the study area for which data on precipitation quality is available. The mean monthly concentrations of Ca^{++}, Mg^{++}, Na^{+}, K^{+}, SO_4^{2-}, and Cl^{-} for the monsoon season have been reported by Maske & Nand (1982) for Srinagar station and by Handa et al. (1984) for Lucknow station. Since Nainital is located nearly half-way between these two stations the mean of the values recorded at Srinagar and Lucknow should fairly represent the precipitation quality at Nainital. The average concentrations of the ions in precipitation during the monsoon season (period of maximum precipitation in the study area) in mg/L is then Ca^{2+} 4.44, Na^{+} 1.02, K^{+} 0.51, and Cl^{-} 0.99.

3.2.6 Stable isotope ratios

The importance of stable isotope ratios in precipitation (δp), the altitudinal variation and the relation between δp and local meteorological conditions have been already presented in Section 2.4.7. In the following sections the stable isotope characteristics of precipitation in the Lake Nainital basin have been presented.

3.2.6.1 Materials and methods

The precipitation samples were collected from four stations set up at different altitudes (Table 3.4), among which three were set-up within the basin and one on the downstream side of the lake basin (Figure 3.2). The samples were collected 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 representative integrated sample. All necessary precautions were taken to avoid evaporative enrichment during collection and storage. The collected samples were analysed by Stable Isotope Ratio Mass Spectrometer at the Isotope Division, Bhabha Atomic Research Centre, Mumbai, India. The samples were analysed using the CO_2 equilibration method for $\delta^{18}O$ (Epstein and Mayeda, 1953) and by passing through zinc shots for δD (Coleman et al., 1982). The isotope analyses were carried out in two independent VG Isogas (model: 602E) mass spectrometers dedicated to the oxygen and hydrogen isotope ratio measurements. Navada and Kulkarni (1989) have presented in detail, the laboratory procedure followed for the stable isotope analyses. The measurement precision for $\delta^{18}O$ was ±0.1 ‰ and for δD was ±1‰. All the isotope data reported in this article correspond to VSMOW.

The meteorological data such as temperature, relative humidity and rainfall were observed at the study area during the period 1994-95 and the same have been used in the statistical analyses. The amount of precipitation was calculated using weighted average of the data obtained from three self recording (siphon type) raingauges and one manual raingauge installed within the basin (Figure 3.2) by Theissen polygon method.

3.2.6.2 Local meteoric water line (LMWL)

Meteoric water line is the best fit line of the oxygen-18 and deuterium content of the fresh waters. Craig (1961) after a global survey of stable isotope contents in freshwaters proposed a Global Meteoric Water Line (GMWL), which was later modified by Rozanski et al. (1993). The equation (2.11) for the GMWL on the basis of exhaustive isotope data collected through the IAEA/WMO world wide network is given by Rozanski et al., 1993. The GMWL is essentially a global average of several Local Meteoric Water Lines (LMWL). The LMWL is controlled by the local climatic conditions and the source of the vapour mass, particularly the slope of the line is influenced by the secondary evaporation. The knowledge of LMWL is essential for regional or local hydrological studies.

The statistical analyses using the isotopic data of the Nainital precipitation (Table 3.4) yielded the following equation for the local meteoric water line for the study area:

$$\begin{gathered}\delta D\ \text{‰} = 7.5 \cdot \delta^{18}O + 4.82 \\ (n = 15;\ r = 0.97)\end{gathered} \tag{3.3}$$

Equation 3.3 represents precipitation during the monsoon period i.e. during the months of July, August and September. The equation compares very well with that proposed by other investigators (Bhattacharya et al., 1985; Seigel and Jenkins 1987; Krishnamurthy and Bhattacharya, 1991; and Bartarya et al. 1995). Krishnamurthy and Bhattacharya (1991) analysed a relatively large database on groundwater samples collected from northern India and proposed a regional meteoric water line with a slope of ~7.2 as a characteristic of Indian monsoon. Bartarya et al. (1995) proposed an equation with a slope of ~7.1 and intercept of ~15 for Gaula basin of the Kumaun Himalayas. However, the equation by Bartarya et al. (1995) is not based on isotopic composition of monthly-integrated precipitation samples, but on randomly selected individual storm events.

When Equations (2.11) and (3.3) are compared the change in slope as well as intercept can be explained, only if the effect of secondary evaporation during rainfall is considered. Bhattacharya et al. (1985), Krishnamurthy and Bhattacharya (1991) and Datta et al. (1991) have also reported the effect of secondary evaporation in Indian precipitation. Ehhalt et al. (1963) reported that under dry conditions the re-evaporation of falling rain drops has an effect of obscuring the seasonal variations. But the secondary evaporation of falling raindrops has similar effects in the present case where the moist climatic conditions prevail during the study period. Further, this re-evaporation overshadows the true altitude effect as discussed in what follows.

Table 3.4 The isotopic composition of precipitation samples collected at different altitudes in the Nainital basin and downstream area during months indicated. Elevations given are with reference to mean sea level.

Site	Elevation, m	Month	$\delta^{18}O$, ‰	δD, ‰	'd', ‰
Snow View	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 the LMWL.

3.2.6.3 'd' Excess Parameter

In $\delta^{18}O$-δD plot when a line with a slope equal to 8 is drawn through the data points, the intercept 'd' (also called d-excess parameter) on the δD axis indicates the effect of secondary evaporation. The 'd' values at different altitudes for precipitation during any given month should be comparable, if secondary evaporation is not significant. In other words the 'd' values should have poor correlation with the altitude. While this is true for the study area for the July and August precipitation, a significant correlation is observed for the month of September (Table

3.5). This secondary evaporation gives rise to 'secondary-altitude effect' in mountainous regions (Coplen, 1993). This 'secondary-altitude effect' is different from the 'altitude effect' which is a direct resultant of the adiabatic cooling related rain-out process as originally perceived by Dansgaard (1964).

Table 3.5. Estimated altitude effects (per 100 m) in $\delta^{18}O$ and δD for the monsoon precipitation of 1995 at Nainital, Kumaun Himalayas.

Month	$\delta^{18}O$, ‰	δD, ‰	'd', ‰
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

3.2.6.4 Altitude effect in stable isotope ratios

The precipitation data for the monsoon season of 1995 (Table 3.4) have been used in the estimation of true mean altitude effect. The elevation versus isotopic ratios or δ values of precipitation (δp) plots are given in Figure 3.7a and 3.7b. The estimated altitude effects for different months in $\delta^{18}O$ and in δD are presented in Table 3.5. It is observed that the $\delta^{18}O$ data shows a higher slope for September than for July or August. As discussed in the preceding section, this higher slope is 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 the air (Table 3.5). On the other hand, the δD data of September does not show this effect. This is because in case of 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. Literatures available on precipitation isotope characteristics cite altitude effect mainly as a resultant of adiabatic cooling and related rain-out process. Thus, the rainfall weighted mean altitude effect in $\delta^{18}O$ is estimated using the isotopic data of July, August, and September, 1995 and is:

Figure 3.7a Altitude effect in del O-18 during the monsoon season of 1995 based on the rainfall weighted averages.

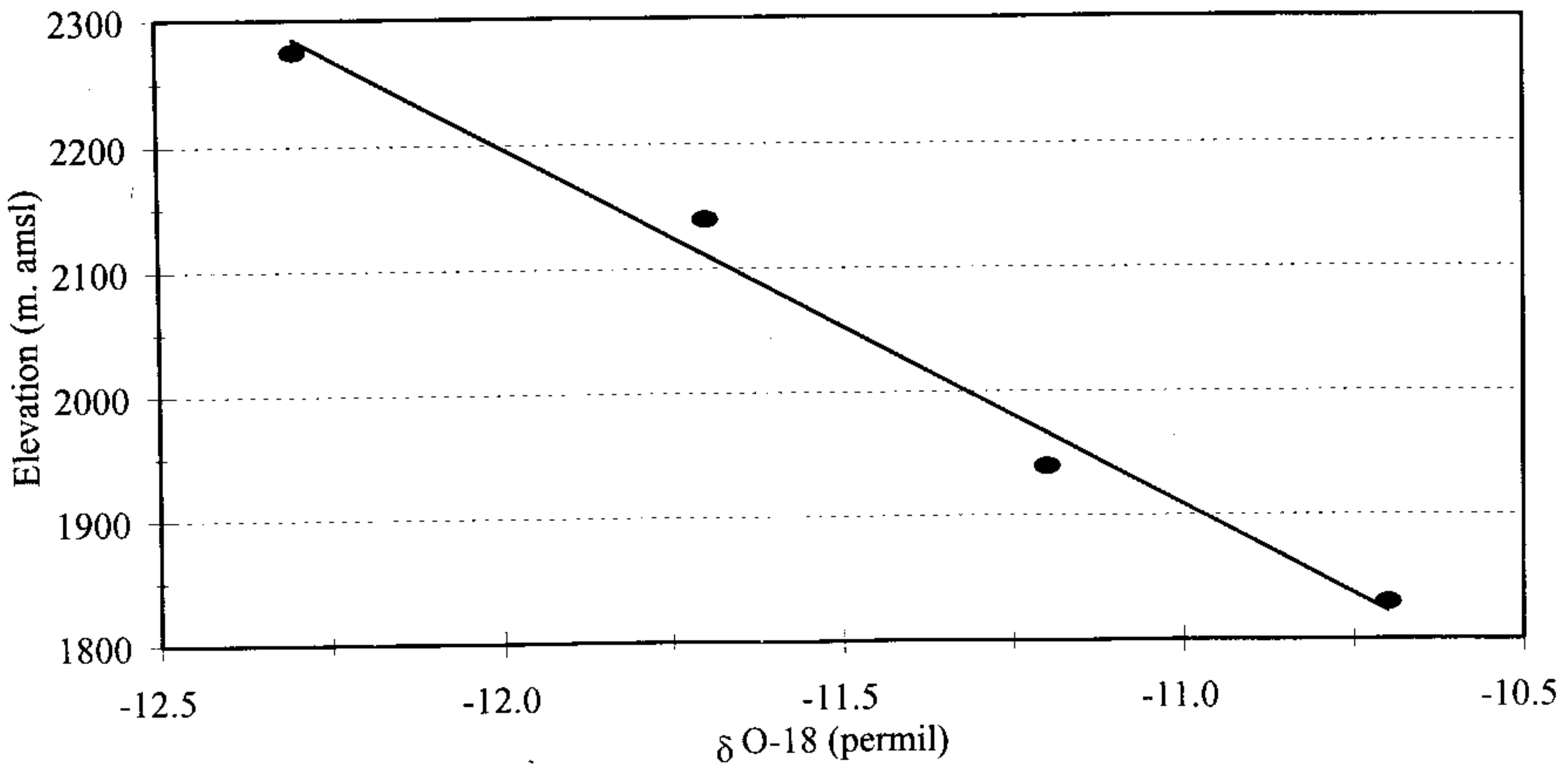
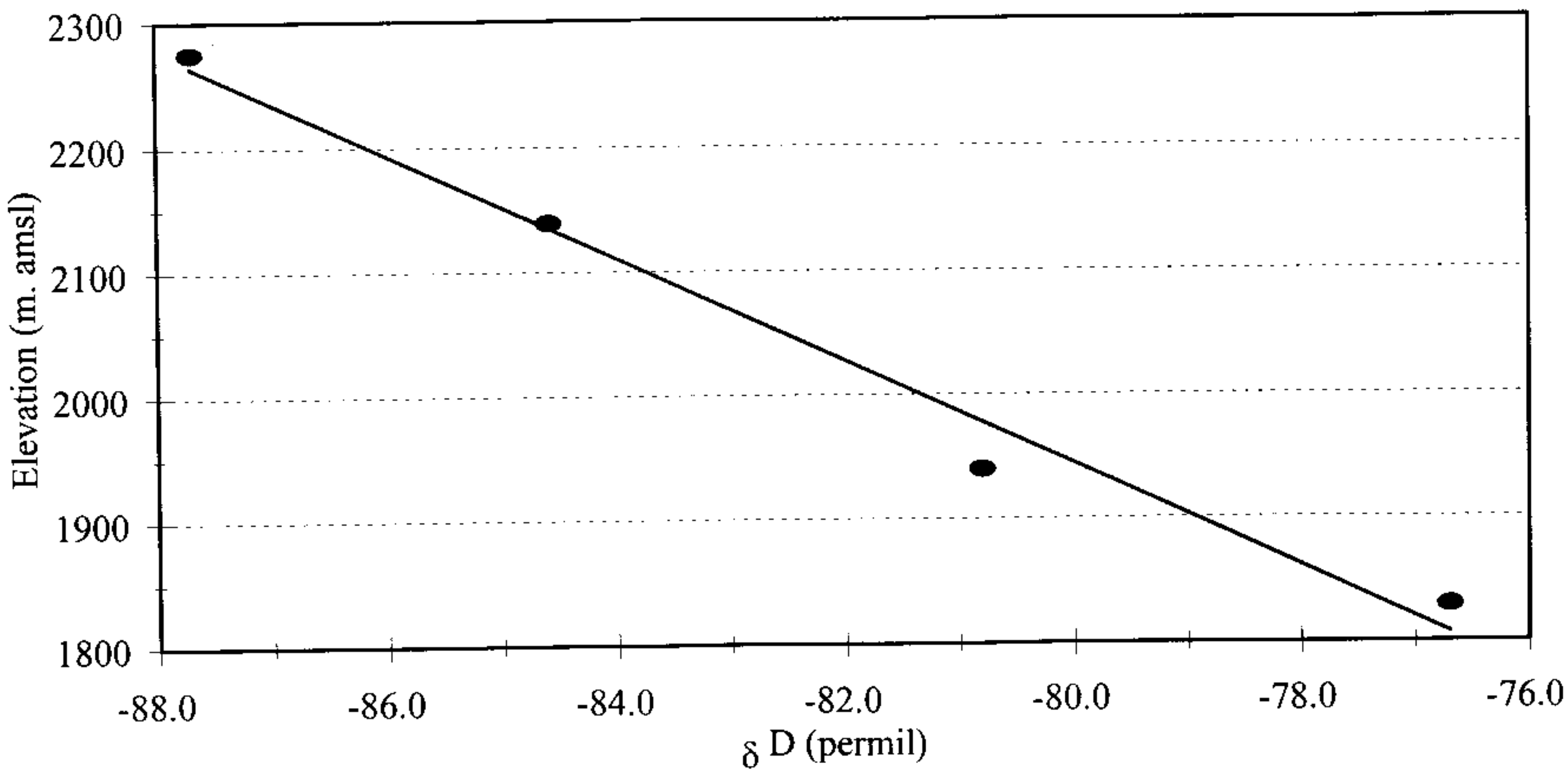


Figure 3.7b Altitude effect in del D during the monsoon season of 1995 based on the rainfall weighted averages.



$$-0.34 \pm 0.10‰ / 100m \; [r=0.99]$$

The rainfall weighted mean altitude effect in δD for the same months is:

$$-2.4‰ \pm 0.8‰ / 100m \; [r=0.98]$$

The mean altitude effects presented above are within the range of values reported from other parts of the world (Yurtsever and Gat, 1981; Clark and Fritz, 1997). However, the values are higher than -0.19‰ for $\delta^{18}O$ and -1.6‰ for δD derived by Ramesh and Sarin (1992) through isotopic compositions of headwaters of river Ganges from the Himalayas. The lower values of Ramesh and Sarin (1992) can be ascribed to the fact that the river at different points of its reach, represents the integrated run-off up to the respective points. In addition, the contribution from groundwater that is recharged at different altitudes will also have significant dampening effect on the isotopic signatures of the river waters. The process of run-off also gives rise to evaporative enrichment of the isotope signatures of the river waters.

The mean altitude effect observed in the present study are also significantly higher than the altitude effect reported for $\delta^{18}O$ (0.14‰ per 100 m) by Bartarya et al. (1995) who used the spring water isotopic compositions. This low value is due to the difference in altitude of recharge areas of springs and their outlets. This fact clearly brings out the risks in using surface expressions of subterranean waters as proxy to study the altitude effect in δp.

3.2.6.5 Local meteorological conditions and δp

The δp data if not available for a particular location, could be generated using the multiple / linear regression equations developed using data from the vast IAEA network. Yurtsever and Gat (1981) have presented equations with different independent variables viz., latitude, temperature and amount of precipitation. However, they pointed out that this approach may be true on a global scale and may not hold good on a regional scale and suggested that certain geographical / climatological parameters may be included for better results.

In the present case δp data are generated for the period for which they are not available, by using available δp data on the location itself. For the purpose, the isotopic data collected at the lake site (Table 3.6) have been used for the regression analyses. Two different combinations of independent variables are tested, viz. a) monthly mean temperature (T_m), monthly mean vapour pressure (EA_m) and monthly rainfall (RF) and b) T_m, RF, and monthly mean relative humidity (RH_m). The first combination gave good results with a multiple correlation coefficient of 0.76 and a standard error of 2.6‰, while the latter combination gave even better results with a multiple correlation coefficient of 0.88 and a standard error of 2.04‰. The equation developed is:

$$\delta^{18}O‰ = (-0.30)\ T_m + (-0.27)\ RH_m + (0.01) RF + 8.98 \quad (n = 10;\ R = 0.88) \tag{3.4}$$

Where, T_m is in °C, RH_m in % and RF in mm. The above equation has been used to generate the generate the stable isotope data for the months for which the data are not available during the study period.

Table 3.6. Isotopic composition precipitation and the monthly meteorological data of the corresponding month of sampling at the lake site.

Month of sampling	$\delta^{18}O$, ‰	δD, ‰	T_m,°C	RH_m,%	RF, mm
Sep., 1994	-12.6	-88	17.7	68.4	43
Feb., 1995	-7.6	-	6.4	58.6	59
Apr., 1995	-1.6	-7	15.3	32.6	6
May, 1995	-10.5	-64	20.2	40.9	19
Jul., 1995	-11.4	-79	18.8	82.3	509
Sep., 1995	-10.5	-77	17.2	77.9	341
Aug., 1995	-11.3	-83	18.4	85.5	632

3.2.6.6 Interannual variability in isotope ratios:

The short data length (one season only) do not permit an understanding of the precipitation - catchment hydrology relations using the isotope data. However, it may be attempted by using a comparatively long term data record available from near by stations. The nearest station for which such isotope data available is New Delhi, that is an IAEA/WMO GNIP station. For New Delhi, which is located at a distance of ~300 km south of Nainital, the $\delta^{18}O$ and δD data are available from 1961 to 1995. For the present study only the monsoon precipitation data are considered as maximum amount of rainfall is received only during this season. The equation of the New Delhi MWL is:

$$\delta D = (7.9 \pm 0.3)\ \delta^{18}O + (7.6 \pm 2.0) \qquad (3.5)$$
$$(n = 27;\ r = 0.96)$$

The slope (7.9) of the New Delhi MWL (monsoon season only) is very close to that of GMWL, than when the complete annual record is considered. As discussed in the previous sections it is mainly the non-monsoon precipitation that reduces the slope of the MWL. The interannual variation in the weighted isotopic ratios of the monsoon season precipitation ranges from lighter ($\delta^{18}O$ = -12.0‰, δD = -86‰) rainfall of 1995 monsoon to heavier ($\delta^{18}O$ = -1.6‰, δD = -5‰) rainfall of 1987 monsoon. The 'd' excess parameter also varies from a minimum of 1.0‰ in 1985 to a maximum of 15.2‰ in 1974. Since the moisture source for the monsoon precipitation at both Nainital and New Delhi sites are same, it is possible that the isotope ratios in monsoon precipitation of Nainital may also show interannual variations.

3.3 Hydrology

3.3.1 Landuse Pattern

For the landuse classification of the lake catchment digital data of the SPOT satellite of resolution 20m and 10m, for the period November, 1987 and March, 1989 were processed. The digital satellite imageries were coupled with the base map of Lake Nainital catchment prepared using the Survey of India toposheet #53 O/7 (scale 1:50,000) in a GIS environment using ILWIS® and ERDAS® software packages. Using supervisory classification approach the catchment area was classified into five landuse classes (Table 3.7).

Table 3.7 Landuse data of Lake Nainital basin using remotesensing data

CLASS	Percent area
Water bodies	10.4
Forest	48.4
Barren	18.3
Built-up	19.3
Shadow	3.6

Since no ground truth studies were carried out the above results may at best be considered as preliminary only. Further, the scale of the map and the satellite images did not facilitate detailed classification so as to include minor landuse classes such as roads, play grounds formed over the land slide debris etc. Therefore, the landuse classification of the Lake catchment was also attempted by another conventional method. In the alternative method, the Survey of India Nainital guide map of scale 1:10,000 was used in conjunction with the Forest map of Nainital (scale - 1:7875) obtained from the Uttar Pradesh Forest Department. On site verification (ground truth study) was carried out in the catchment to establish the landuse. The area under different landuse was measured using a Digital Planimeter. The area under roads and pathways were estimated by measuring the length using an Opisometer and assuming uniform width. The results obtained were moderately different from those obtained by the remote sensing - GIS method and are given in Table 3.8.

Since the field verifications were carried out and that the base maps used for the study were of smaller scales, the results obtained by the second method may be considered as more reliable.

Table 3.8 Landuse data of Lake Nainital basin using conventional approach

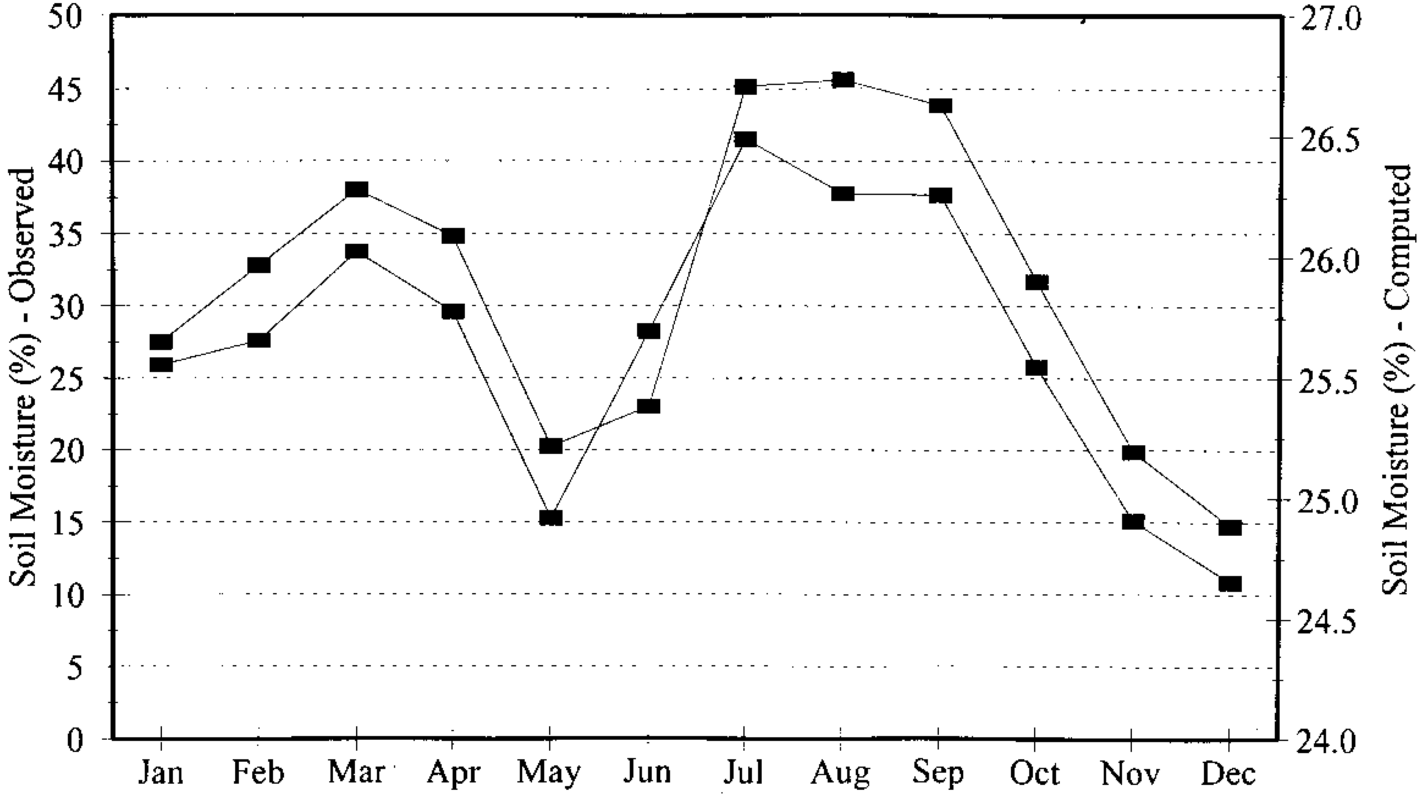
CLASS	Area (km2)	Percent area
Reserved Forest	0.9932	21.12
Other forests and shrubs	0.9689	20.60
Built-up	1.9195	40.81
Roads	0.1000	2.13
Water bodies	0.4858	10.33
Play ground	0.0527	1.12
Barren Land	0.1829	3.89

3.3.2 Soil properties

Soil samples collected from different parts of the lake basin were subjected laboratory investigations to determine the field capacity and wilting point. The pressure plate apparatus (Eijelkamp make) was used for determining the moisture retention characteristics of the soil samples. Saturated samples were placed in the pressure plate apparatus and pressures that correspond to the wilting point (15 bars) and field capacity (0.15 bar) were applied till reaching equilibrium. Samples were then removed from the pressure chambers and weighed on a high precision electronic balance to record the moist weight of the samples. The weighed wet samples were then placed in an oven at 105°C for 24 hours and dried. The dry weight was then recorded. The moisture retention of the soil samples were then determined using the moist and dry weights of the samples. The results of the laboratory investigation are presented in Table 3.9.

The soil moisture variation with time has also been estimated by mass balance approach (Soni, 1987) for soil mositure stoarge estimation by considering the precipitation input and evapotranspiration and runoff losses. The soil moisture thus computed is shown along with the observed soil moisture in Figure 3.8. The difference in the values are due to: a) the computed soil moisture pertains to the dominant soil type in the catchment namely the clay loam topping the calcareous shales and other rock type, while the observations were made at the coarser soil at the

Fig. 3.8 Variation of observed and computed soil moisture in Lake Nainital catchment



Flat in the Mallital bank of the Lake (due to logistical problems the soil moisture sensor could not be located at an apt site in the catchment) and b) the in-situ calibration of the soil moisture sensor could not be carried out, and had to rely on the factory calibrations. However, the pattern of soil moisture variation is very similar in both observed and computed values (Figure 3.8).

Table 3.9 Moisture retention characteristics of soil samples collected from different sites in Lake Nainital catchment.

Site	Field Capacity (0.15 bars)		Wilting Point (15 bars)	
	soil moisture content	Average	soil moisture content	Average
Alma Cottage	0.3122	0.3125	0.1958	0.1997
	0.3093		0.2004	
	0.3159		0.2029	
Ayarpatta - 1	0.2090	0.2035	0.0865	0.0856
	0.1990		0.0846	
	0.2020			
Ayarpatta - 2	0.1952	0.1996	0.1139	0.1146
	0.1987		0.1159	
	0.2049		0.1141	
Staff House	0.3540	0.3489	0.2416	0.2446
	0.3466		0.2391	
	0.3460		0.2532	
Catchment average	F.C.	0.2661	W.P.	0.1611

3.3.3 Infiltration

The infiltration is defined as the downward movement of water through soil surface due to combined forces of the capillarity and gravity contributing to replenishment of soil moisture, recharge to groundwater and ultimately augmenting streamflows during lean seasons. Infiltration rate is the volume of water entering the soil profile per unit surface area and time. Although several factors such as intensity of rainfall, soil characteristics, soil moisture content, vegetal cover, land use, entrapped air and weather etc. influence the infiltration characteristics of a catchment or basin, one of the chief factors is the pore size distribution in the soil column (V. P. Singh, personal communication). In the mountainous terrain as the flow gets concentrated along the valley bottoms, knowledge of infiltration characteristics along the natural surface drainage channels are vital for hydrological studies of mountainous catchments. Keeping the above facts in view, few infiltration tests were conducted in the Lake Nainital catchment.

The infiltration tests were conducted during the month of August, 1998 i.e. during the mid-monsoon season, so that the soil cover will be near saturation. The tests were conducted using single ring infiltrometer of 30 cm inner diameter and 45 cm height. the bottom rim of the ring was bevelled to facilitate easy-drive into the soil with less disturbance. The infiltrometers were installed such that 30 cm of the ring was inside the soil column and 15 remained exposed. Polythene sheets were spread inside the infiltrometer prior to the actual start of the test. The sheets were removed instantly so that the initial infiltration rate could be obtained. However, as the soil was fully saturated no decreasing trend was seen in the infiltration rate during the test period in all the three test sites. A constant head of 5 cm was maintained during the tests with the help of a storage tank, water level controller and solenoid valve set-up developed at the National Institute of Hydrology. The tests were conducted at three different sites in the Lake Nainital catchment (Figure 3.2), they are:

i) Assembly Cottage campus near the Lakeview Cottage along the Birla Vidyamandir Road. The site was characterised by grassy land, slope, macropore dominated weathered shale with moderate soil cover. The final infiltration rate was ~50 cm/hr.

ii) Forest Chauki along the Birla Vidyamandir Road. The site was characterised by thick clayey soil. The final infiltration rate was ~1.1 cm/hr.

iii) Sukhatal lake bed. The site was characterised by flat land, macropore dominated, shattered and weathered shale with little soil cover. The final infiltration rate was ~78 cm/hr.

From the above results it is seen that the infiltration rates in the Lake basin vary greatly. The higher infiltration rates along the valley bottoms could be a reason for the low rainfall - runoff relationship (25-30%) in the Lake Nainital catchment (cf. section 3.3.6). Further, the higher infiltration rate in the Sukhatal lake bed could be the reason for the drying-up of the Lake Sukhatal at the end of monsoon season.

3.3.4 Evapotranspiration

3.3.4.1 Potential Evapotranspiration

Conceptually potential evapotranspiration (PET) may be defined as the rate at which evapotranspiration would occur from a large area completely and uniformly covered with growing vegetation that has access to an unlimited supply of soil water without advection or heat storage effects. Although PET depends largely on climate, the surface and vegetative characteristics have a strong influence on PET. This has lead to the term Reference ET, rather than PET in some recent literatures, as basically ET is estimated using a particluar vegetation as a reference. In hydrological studies the methods used for the estimation of PET can be broadly classified into Temperature methods, radiation methods, combination methods and pan methods. Out of these four the combination methods are mostly preferred, if the required data are available.

The potential evapotranspiration in the Nainital catchment has been computed by the following combination and temperature methods:

1. Penman
2. Christiansen
3. Makkink
4. Jensen & Haise
5. Thorntwaite
6. Blaney-Criddle (SCS)

The results are presented in Table 3.10 and for comparison shown in Figure 3.9. It is seen that the PET estimated by Penman method is higher than the estimates made by other methods. The Pan evaporimeter results are also shown in Figure 3.9.

3.3.4.2 Actual Evapotranspiration

One of the most widely used approaches for estimating the actual evapotranspiration makes use of the potential evapotranspiration (estimated using meteorological data) and the relative soil water content. This approach has been adopted in this study. The actual evapotranspiration was computed using the following equation (Dingman, 1993).

$$ET_A = ET_P \cdot \left(\frac{\theta - AWP}{AFC - AWP} \right) \qquad (3.6)$$

Where ET_A = Actual evapotranspiration (mm)

ET_P = Potential evapotranspiration (mm)

θ = Soil moisture content

AWP= Average Wilting Point

AFC = Average Field Capacity

The actual evapotranspiration calculated using the PET values are shown in Figure 3.10.

Fig. 3.9 Variation of Potential evapotranspiration (penman method) for Lake Nainital basin. Data pertaining to the year 1995.

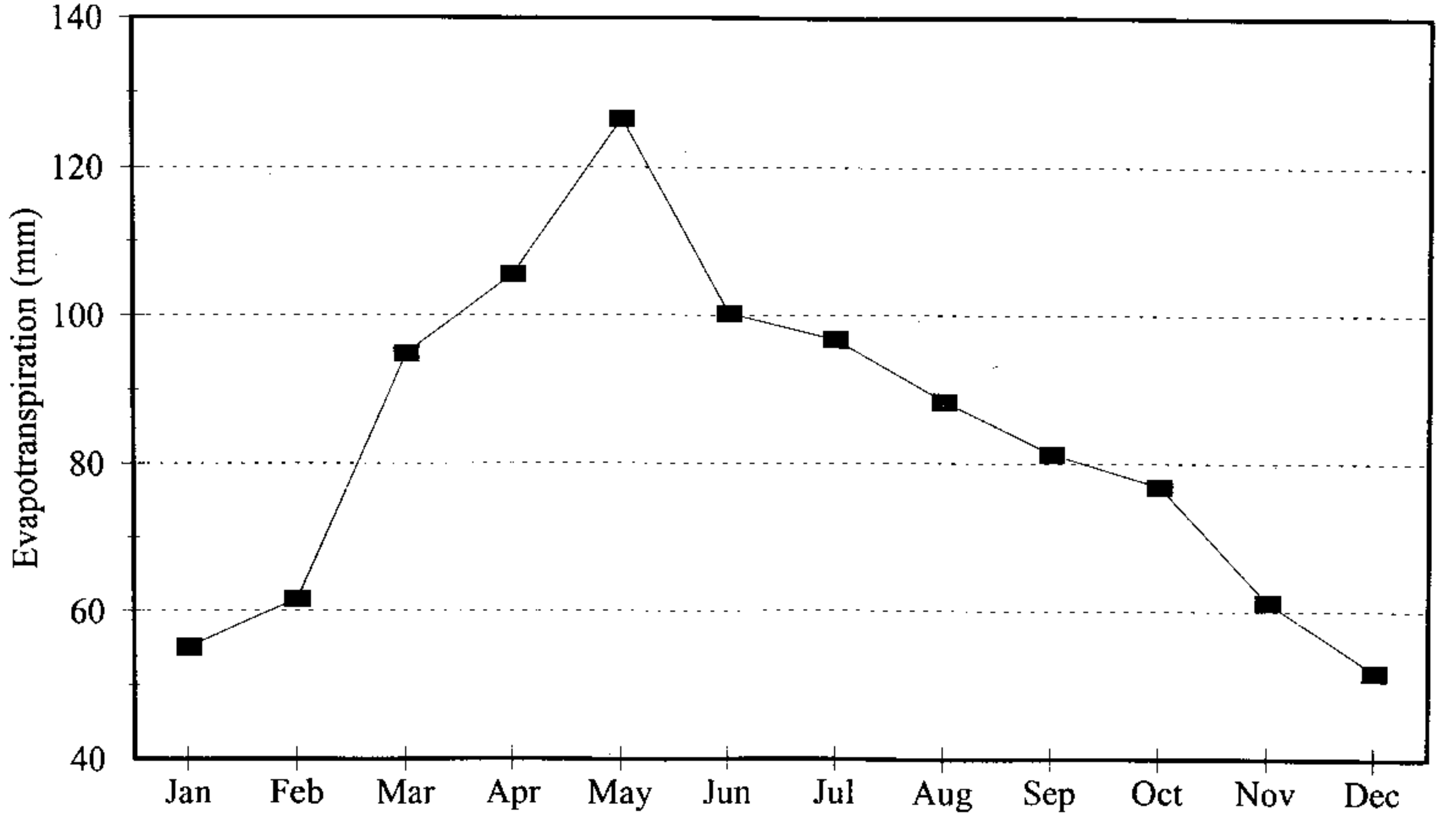
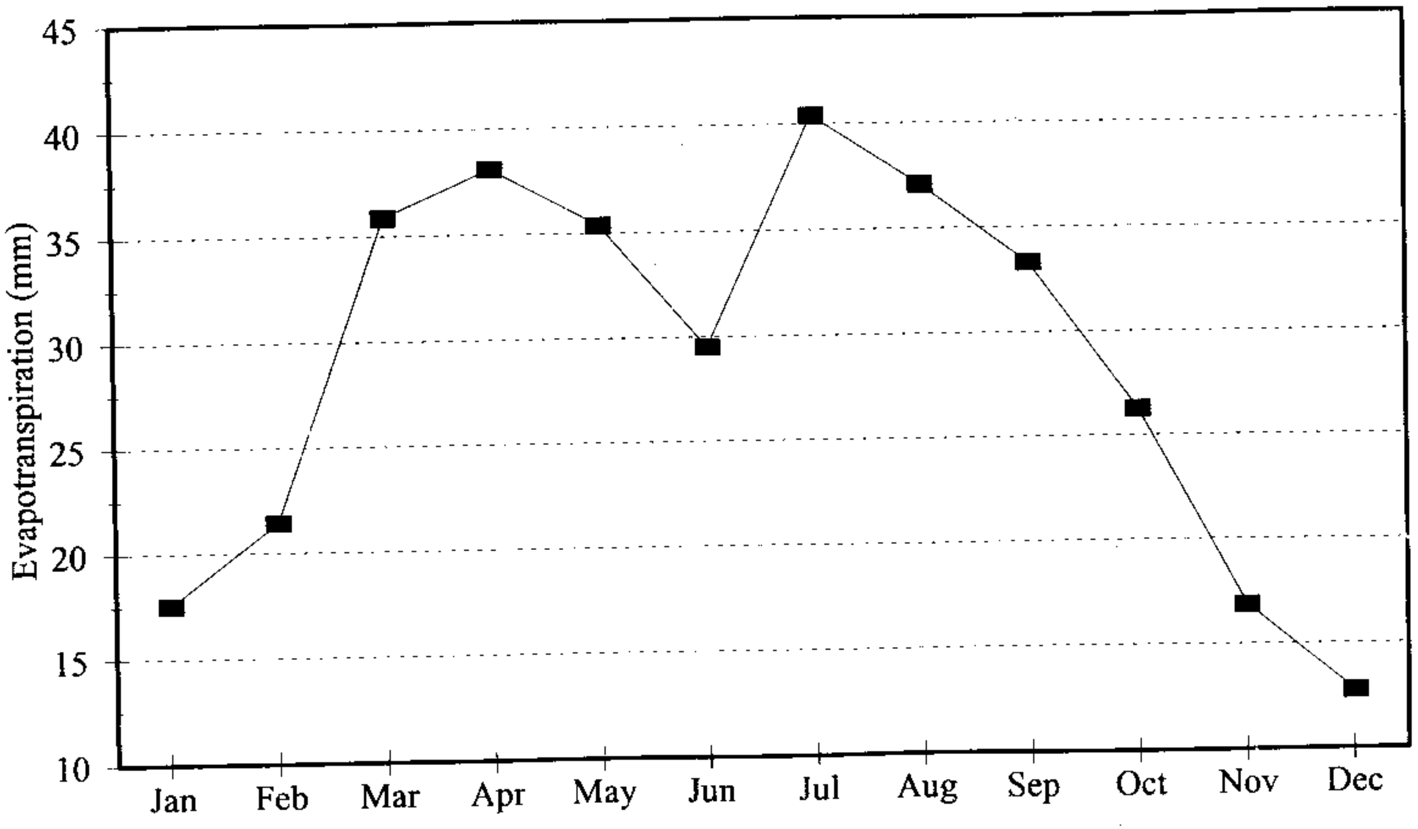


Fig. 3.10 Variation of Actual evapotranspiration (soil moisture retention method) for Lake Nainital basin. Data pertaining to the year 1995.



3.3.5 Drainage

The drainage pattern in the Nainital lake basin is controlled by the structural conditions. The drainage map of Lake Nainital (Figure 3.2) basin clearly brings out the influence of geology, as the area to the west (Ayarpatta ridge) of the Lake is marked by very few streams while the area to the east (Sher-ka-Danda ridge) is characterised by several parallel to subparallel streams. Rawat (1988) ascribed this parallel drainage in the Sher-ka-Danda ridge to the presence of several fractures, faults and joints, and the absence of a dense drainage network in the Ayarpatta to the presence of dolomites and limestone of Upper Krol. This fact is also reflected in the fluvial morphometric parameters of Lake basin given in Table 3.11.

Compared to the Shales and marls of Lower and Middle Krol formations, the Dolomites and limestones of Upper Krol exhibit a lower values of the three key morphometric parameters given above. There are about 20 channels that drain the entire Lake basin. Almost all these channels were partially or fully lined during the early part of the century and all the channels join the Lake Nainital. Out of these 20, only two are perennial, i.e. with some flow through out the year. Others are seasonal and flow only after rainfall events. The perennial ones are a) the Nainadevi drain (Drain # 23) whose flow is sustained mainly by discharge from Pardhadhara spring (located near Nainital Water Works) and sewerage disposal from some part of the residential area located on the banks of the channel and b) the Rickshaw stand drain whose flow is sustained mainly by sewerage disposal.

Table 3.11 Fluvial morphometric parameters of Lake Nainital basin (After Rawat, 1988)

Geologic formation	Rock type	Morphometric parameters		
		Drainage density (km/km^2)	Stream frequency (# of channels/km^2)	Bifurcation ratio
L. Krol	Shales & Marls	6.05	8.15	4.50
M. Krol	Red Shales & Marls	4.15	5.30	3.20
U. Krol	Dolomites & Limestones	0.30	1.25	2.00

3.3.5.1 Flow through the drains

As noted in the previous section, most of the drains are generally dry and flow only after rainfall events. However, the Nainadevi drain and Rickshaw stand drain flow through out the year. The discharge in these two drains were measured using a pygmy current meter and the flow rates were estimated using the velocity-area method. Weekly measurements were made in these two perennial drains, assuming that the flow in the drains do not vary significantly during the intermittent period. The discharge of the other drains that flow on non-rainy days during monsoon season due to interflow and/or groundwater flow that join the drains, were measured on non-rainy days of monsoon season only.

3.3.6 Chemistry of drain waters

Water samples were collected from the drains for chemical analyses by grab-sampling method near the drain axis. All the samples that were collected from the drains were preserved and stored as per the procedure presented in section 3.4.5.1. The procedures followed for the chemical analyses have also been given in section 3.4.5.1. The number of samples collected from the drains during different sampling campaigns in the study period are given in Table 3.12.

Table 3.12 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	Dec.,95	-
Feb.,95	1	Jan.,96	2
Mar.,95	2	Apr.,96	2
May, 95	2	Jul.,96	1
Jun.,95	-		

Samples were collected from Nainadevi Temple drain and Rickshaw stand drain during almost all the campaigns, while samples from other drains, particularly those flowing in the Sher-ka-Danda hill were collected only during monsoon months. In general, in all the drain waters the concentration of cations were found in the following order $Mg^{2+} > Ca^{2+} > Na^{+} > K^{+}$ and in case of anions $HCO_3^- > SO_4^{2-} > Cl^-$. According to Stiff's classification the drain waters may be classified as Magnesium Bicarbonate type. The observed range and mean concentrations of different anions/cations along with the corresponding statistical errors in the mean of the samples collected during non-monsoon season are given in Table 3.13.

From the data it is seen that there is a significant variability in the concentration of different ions. The variation is mainly due to seasonal effect and that the higher values are observed during pre-monsoon season. The variability is more pronounced for ions of lower concentration viz., Na, K and Cl, indicating that contamination due to sewerage and domestic waste disposal into the drains. Larger flows in the drains are observed in monsoon season, and that the chemistry of the drains during this season will be influencing the chemistry of the lake.

Table 3.13. Chemical characteristics of drain waters during non-monsoon season

Parameter (mg/L)	No. of samples	Range		Mean	Standard deviation (σ)
		Minimum	Maximum		
Ca^{2+}	14	20	92	64	21.3
Mg^{2+}	14	60	99	75	11.5
Na^{+}	11	8	72	25	18.1
K^{+}	11	6	35	12	8.1
HCO_3^-	13	202	392	313	53
SO_4^{2-}	12	95	194	141	21
Cl^-	14	3	82	31	21
E.C (μS/cm)	10	460	1550	920	300

Therefore, to understand the chemistry of the drain waters during monsoon season all the major drains were monitored and the brief results are given in Table 3.14.

Table 3.14. Chemical characteristics of drain waters during monsoon season

Parameter (mg/L)	No. of samples	Range		Mean	Standard deviation (σ)
		Minimum	Maximum		
Ca^{2+}	20	15	87	42	16.2
Mg^{2+}	20	39	110	74	17.8
Na^{+}	12	12	37	25	7
K^{+}	12	4	22	10	4.8
HCO_3^-	20	140	362	300	47.2
SO_4^{2-}	20	45	199	133	35.7
Cl^-	18	6	46	24	11.2
E.C (μS/cm)	8	480	920	770	135

3.3.7 Stable isotope ratios

A total of eleven samples were collected from the drain waters during the study period for stable isotope analyses, of which four represent the perennial Nainadevi Temple drain and another four the rickshaw stand drain. The remaining three samples represent the seasonal drains viz., Boatclub drain and Library drain. The brief results are given in Table 3.15.

Table 3.15. Isotopic characteristics of drain waters

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	-

The observed values 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 heavier than the local precipitation indicating that some amount of evaporative enrichment could have taken place. The slope of the $\delta^{18}O$ - δD line is ~5.4, that is much less than that of LMWL (Equation 3.3). This also indicates that the drain waters may have suffered evaporative enrichment.

3.3.8 Surface runoff

The surface runoff generated in response to rainfall events has been estimated using two different methods viz. Lake Level Trend Analysis Method and Soil Conservation Service - Curve Number (SCS-CN) Method. The details are described in the following sub-sections.

3.3.8.1 Lake Level Trend Analysis Method

The surface runoff in response to storm events, that reach the lake can be computed from the change in lake storage taking into account the lake level trend. In general, the lake level follow a decreasing trend upto 2nd or 3rd week of June, depending upon the onset of monsoon, beyond which the level starts increasing upto mid September. However, the fluctuations can also be seen in the water levels during rising trend also due to opening of the sluices to drain out the excess amount of water i. e., to regulate the lake level (the maximum lake level is 12.5 feet (=3.81m) in the PWD gauge installed at Tallital bank). The trend of change in the lake level can be defined as the rate of change of water level with respect to input to and output from the lake at any time or time period. The drop in water level indicates lesser input and higher output and on conversely the rise in water level indicates higher input and lesser output from a water body. In addition to this, water level trend (real time change in water level) can distinguish the role of any unknown process taking place. For example, surface runoff can be determined corresponding to a rainfall event if we have the information of trend of change in water level prior to the related inflow. Similarly, if surface runoff is known, the subsurface inflow or subsurface outflow can also determined with the information of trend of change in water level.

The Lake Level Trend Analysis (LLTA) method is not very different from the well hydrograph analyses method used in the estimation of groundwater recharge. The trend represents and accounts for different losses from the lake and also accounts for any subsurface inflow as well as subsurface outflow - as it does not vary drastically over the time interval (24-hours) that is being considered for the estimation of surface runoff. The change in water level in mm over a week's period is averaged and considered as constant for that particular week. Figures 3.11 and 3.12 show the variation in daily lake level during the years 1994 and 1995, respectively, along with daily rainfall during the corresponding years. The equation used for the estimation of surface runoff by LLTA Method is as follows:

$$RO_i = \frac{\left[\frac{(OF_i - T_i - DRF_i)}{1000} * LA_i + O_i\right]}{CA} \qquad (3.7)$$

Where, RO_i = Estimated runoff (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 lake 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
Subscript i = time period (24 hours ending at 0830 hours)

The daily surface inflow estimated by LLTA method has been shown in Figure 3.13a along with that estimated by SCS-CN method (described in the following section) for the year 1995. The monthly estimates of surface runoff by both the methods for the year 1995 are shown in Figure 3.13b. The rainfall - runoff relationship computed using the surface runoff data estimated using LLTA method for the monsoon and non-monsoon seasons of 1995 are shown in Figures 3.14a and 3.14b.

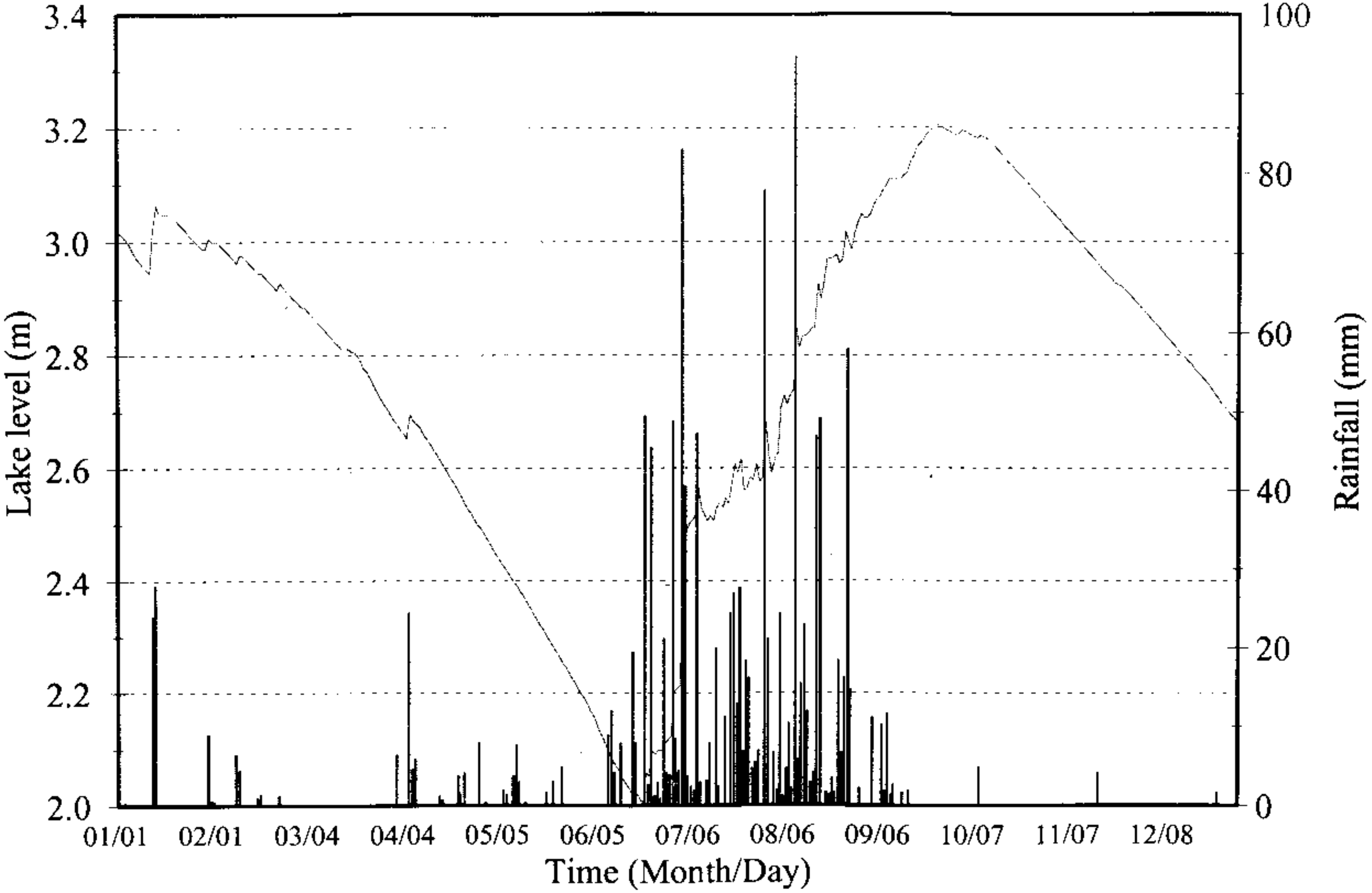


Figure 3.11 Variation of the lake level with daily rainfall during 1994.

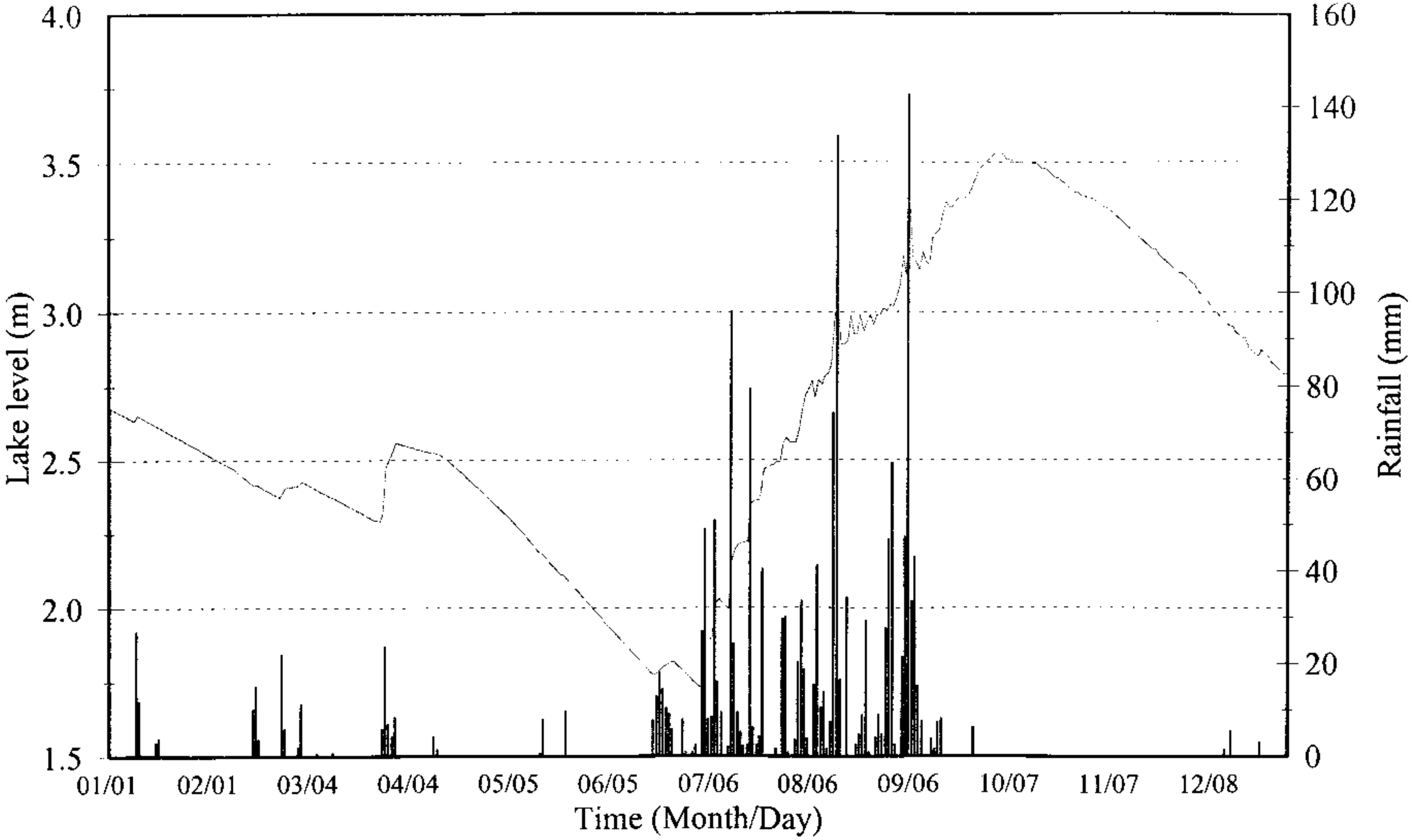


Figure 3.12 Variation of the lake level with daily rainfall during 1995.

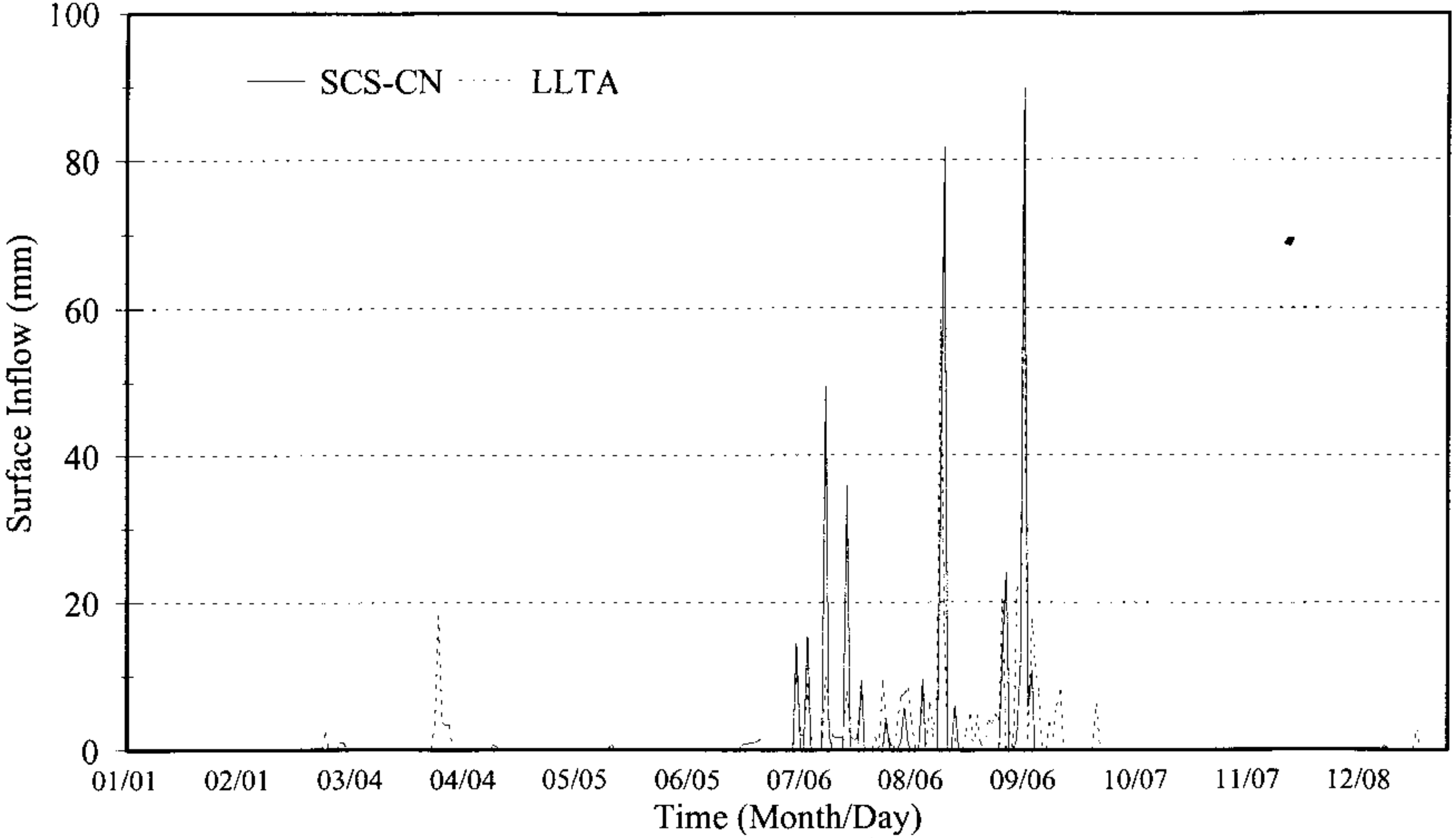


Figure 3.13a Comparison of daily variation in surface inflow to the lake in 1995 estimated by LLTA and SCS-CN methods

Fig. 3.13b Monthly variation in the surface inflow to the lake in 1995 estimated by LLTA and SCS-CN methods

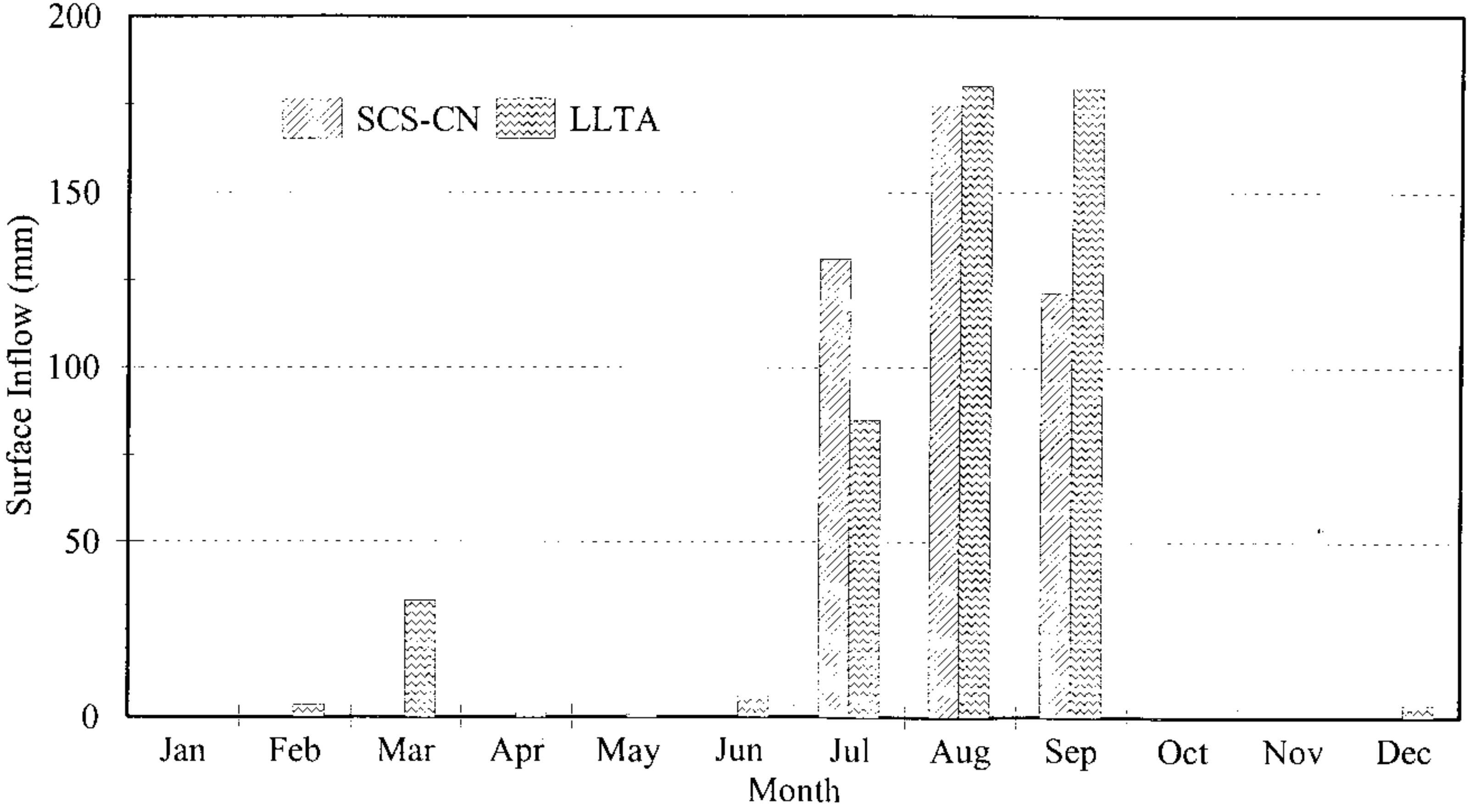


Fig. 3.14a Monsoon season rainfall - runoff relationship in Lake Nainital catchment in 1995 by LLTA method

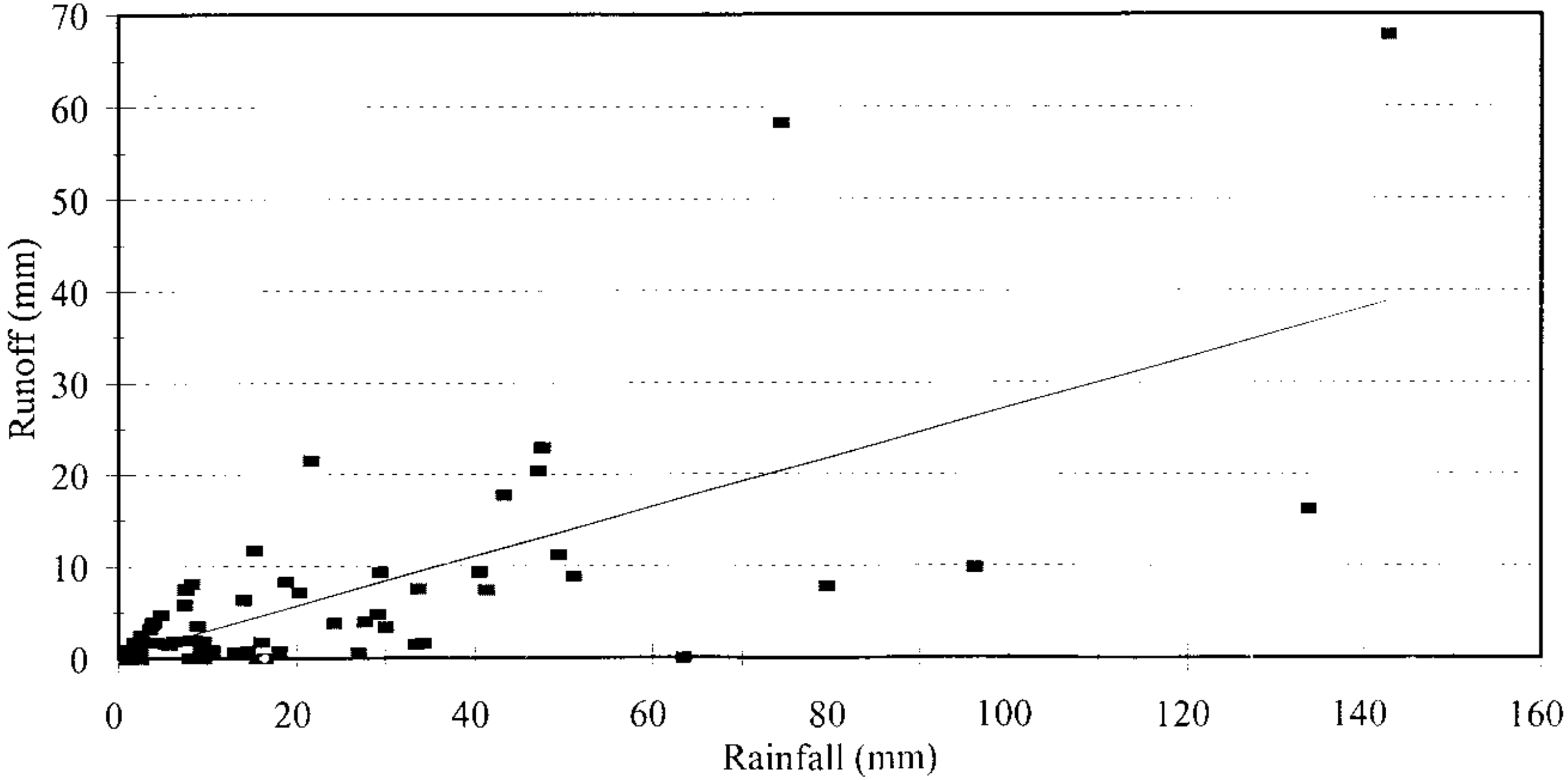
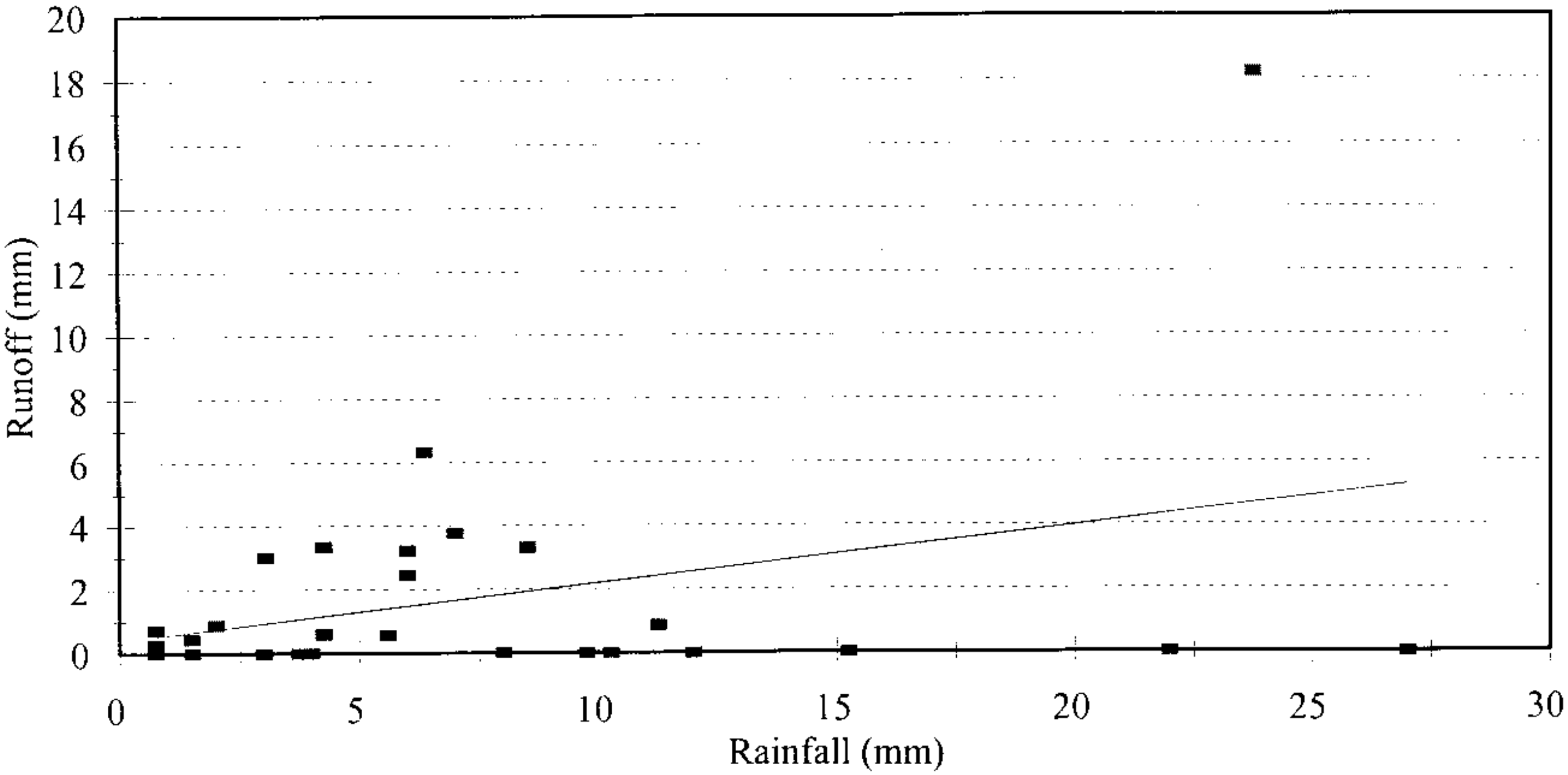


Fig. 3.14b Non-monsoon season rainfall - runoff relationship in Lake Nainital catchment in 1995 by LLTA method



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3.3.8.2 Soil Conservation Service - Curve Number Method

The Surface runoff has also been estimated by Soil Conservation Service - Curve Number (SCS-CN) method. The runoff curve number method has been developed by Soil Conservation Service of US Department of Agriculture. SCS-CN method is more useful for estimation of 24 hour run-off from small catchments, as it is based on 24-hr rainfall-runoff data. In this method, runoff depth is a function of total rainfall depth and an abstraction parameter called curve number (CN). The CN varies from 1 to 100 and it is a function of the following catchment characteristics:

1. Hydrologic soil type
2. Land use and treatment
3. Ground surface conditions and
4. Antecedent moisture condition (AMC).

The basic theory of this well established and widely used method is that the runoff starts after an initial abstraction (mainly interception, infiltration and surface detention) of I_a and that the ratio between actual retention to the potential retention (S) is equal to the ratio of direct runoff (Q) to rainfall (P) minus initial abstraction. i.e.,

$$\frac{P - I_a - Q}{S} = \frac{Q}{P - I_a} \qquad (3.8)$$

rewriting the above equation we get:

$$Q = \frac{(P - I_a)^2}{P - I_a + S} \qquad (3.9)$$

The quantity I_a is related to S so that, $I_a = 0.2$ S. This relation was expressed by Soil Conservation Service based on rainfall-runoff data from small experimental watersheds. Using this relationship, between initial and potential abstractions, equation (3) could be rewritten as:

$$Q = \frac{(P - 0.2S)^2}{P + 0.8S} \qquad (3.10)$$

The parameter S depends upon catchment characteristics as outlined above. The Soil Conservation Service expressed S as a function of Curve Number (CN):

$$S = \frac{1000}{CN} - 10 \qquad (3.11)$$

CN is determined from the hydrologic soil type and antecedent moisture conditions. The hydrologic soil group are labelled as A, B, C or D. A short description of these soil groups are given in Table 3.16.

The antecedent moisture condition (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 the 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. The values that are generally adopted as a measure to determine the AMC level is given in Table 3.17.

Table 3.16. Hydrologic soil classification used in SCS-CN method.

Soil group	Soil characteristics	Minimum Infiltration rate (in./h)
A	Deep sand, deep loess and aggregated silts; High infiltration rates when wetted thoroughly.	0.30 - 0.45
B	Shallow loess and sandy loam; moderate rate of water transmission.	0.15 - 0.30
C	Clay loams, shallow sandy loam, clayey soil; Slow rate of water transmission.	0.05 - 0.15
D	Plastic clays, and soils that swell on wetting; High runoff potential and slow infiltration rates when wetted thoroughly.	<0.05

Table 3.17. Determination of antecedent moisture conditions based on antecedent rainfall

Antecedent Moisture Condition (AMC)	Total 5-day antecedent rainfall (cm)	
	Dormant season	Growing season
I	Less than 1.3 cm	Less than 3.6 cm
II	1.3 to 2.8 cm	3.6 to 5.3 cm
III	More than 2.8 cm	More than 5.3 cm

The appropriate CN for a variety of landuses, soil treatment or farming practices along with hydrological conditions (state of vegetation growth) has been presented by US-Soil Conservation Service (1969) for AMC II. Using the landuse information presented in Section 3.3.1, the CN has been taken from the table of curve numbers for urban areas and agricultural lands presented by SCS and as given in Ponce (1989). Table 3.18 presents the type and percent area under different Landuse, the appropriate CN and the weighted CN for AMC II for the Nainital lake catchment. The CN thus determined for AMC II has been converted to AMC I and AMC III using the following relations (Hawkins et al., 1985):

$$CN_{I} = \frac{CN_{II}}{2.3 - 0.013\ CN_{II}} \tag{3.12}$$

$$CN_{III} = \frac{CN_{II}}{0.43 + 0.0057\ CN_{II}} \tag{3.13}$$

The surface runoff in the lake catchment has thus been computed with the curve numbers determined on the basis of AMCs and daily rainfall data using Equations (3.10) and (3.11), with a spreadsheet software (QProWin). The estimated runoff in units of depth has been converted to units of volume using the exclusive catchment area of the Lake. Since the Sukhatal sub-catchment do not contribute to the surface inflow to the lake (there is no surface outlet from the Sukhatal catchment) it has been excluded in the estimations.

Table 3.18. Determination of composite Curve Number for AMC II based on landuse data of Lake Nainital basin

Land use	Ratio of area under the given Landuse to 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

3.4 Lake Characteristics

3.4.1 Morphology

The maximum length of the crescent shaped lake is 1466 m, maximum width 466 m, maximum depth 27.3 m and mean depth 18 m. The surface area of lake is 463365 m^2. The lake is bound in the east by the Sher-ka-Danda hill, in the north by the landslide deposit called Flat, and in the east by the Ayarpatta hill. To the south of the lake lies the Balia ravine. Most parts of the western and northern banks of the lake are steep, while the eastern bank is characterised by several deltas formed due to deposition of sediment brought in by the 15 drains that drain the Sher-ka-Danda hill. The largest of the deltas found along the banks of the lake is found near the Nainadevi Temple, and is formed due to the sediment brought in by the Drain #23.

Bathymetric survey of Nainital lake has been reported/carried out by several investigators using conventional (manual) sounding methods (Holland, 1896; Rawat,1987). Bathymetric survey carried out by Hashimi et al. (1993) has resulted in gaining bathymetric data-base of crucial import, both for tectonic interpretation and future comparison & studies. The survey was carried out in November, 1989 by using a highly sensitive echo-sounder, sub-bottom profiler and side-scan sonar and Motorala Miniranger III. Hashimi et al.(1993) presented detailed bathymetric contour (1 meter interval) of Nainital Lake floor (Figure 3.15).

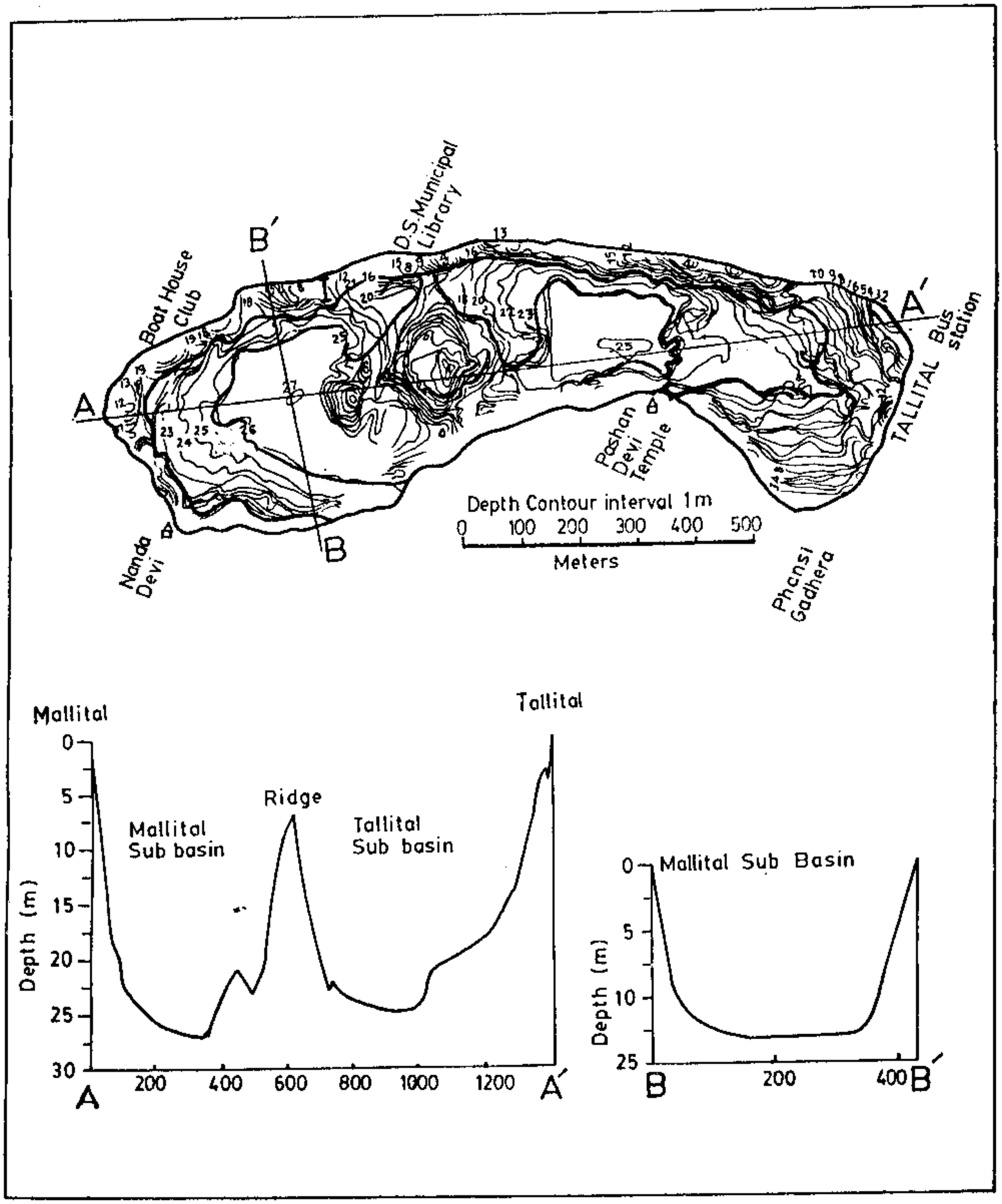


Fig. 3·15 Bathymetric map of Nainital lake (After Hashimi et al., 1993).

The lake is divided into two sub-basins by 100 m wide transverse underwater ridge, 7m to 18m below the lake surface. Although the surface area of the two sub-basins are roughly of the same size, the maximum depth encountered in the northern Mallital sub-basin is ~27.3m and that in the southern sub-basin is ~25.5m. The underwater ridge that divides the lake acts as an effective barrier only below a depth of 18m from lake surface, and therefore a channel like structure facilitates water circulation across the barrier between 7m and 18m depth. Further, a detailed study of the bathymetric contour map of Hashimi et al. (1993) indicates that the dolerite trap found in the southwestern side of the Lake catchment may extend upto the Lake fault that runs through the middle of the lake section 18-3 (Figure 3.15).

3.4.2 Estimation of volume

Information on the lake volume is a basic requirement for any study pertaining to a lake. Computation of the lake volume using depth contours can be made by several methods (Zumberge and Ayers, 1964).

i) Volume of lake may be determined by measuring the area enclosed by the depth contours (the location coordinates plotted against depth) and the area under the curve so obtained may be planimetered or otherwise measured.

ii) In another method, the area enclosed by successive pairs of depth contours is averaged out and multiplied by the contour interval to yield a series of volume elements which are then summed using the following relation:

$$V_{A_1A_2} = \frac{h}{3} * (A_1 + A_2 + \sqrt{A_1A_2}) \qquad (3.14)$$

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

The bathymetric contour map of Hashimi et al.(1993) has been used in the present study to calculate the volume of Lake Nainital. The measurement of area of each contour was carried out by using a digital planimeter. The contours for the shallower depth zone (upto 4 m) were not readable from the map of Hashimi et al. (1993). Therefore, the calculation of area was worked out with the help of Hashimi et al. (1993) map, from 4 m to 27 m depth only. For the calculation of area from surface to 4 m depth, the data reported by United Province PWD (Byrnes, 1929) has been used. Byrnes (1929) has reported the contouring of lake at the interval of 1 foot, from maximum lake level (surface) upto the depth of 12 ft. This data was further interpolated to determine the depth contours in the shallow zone. Thus, the complete area and then the volume of lake, have been computed and are presented in Tables 3.19 and 3.20. The variation of lake area and volume with respect to depth have been shown in Figures 3.16 & 3.17.

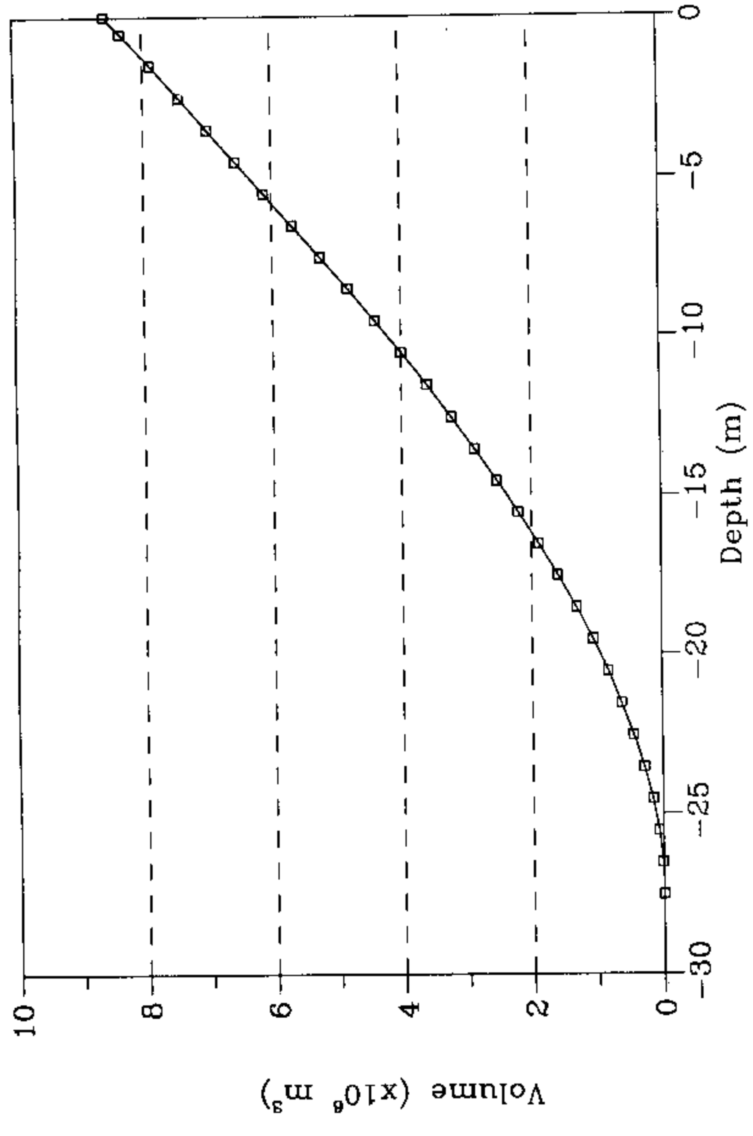


Figure 3.16 Variation of volume of Nainital lake with depth.

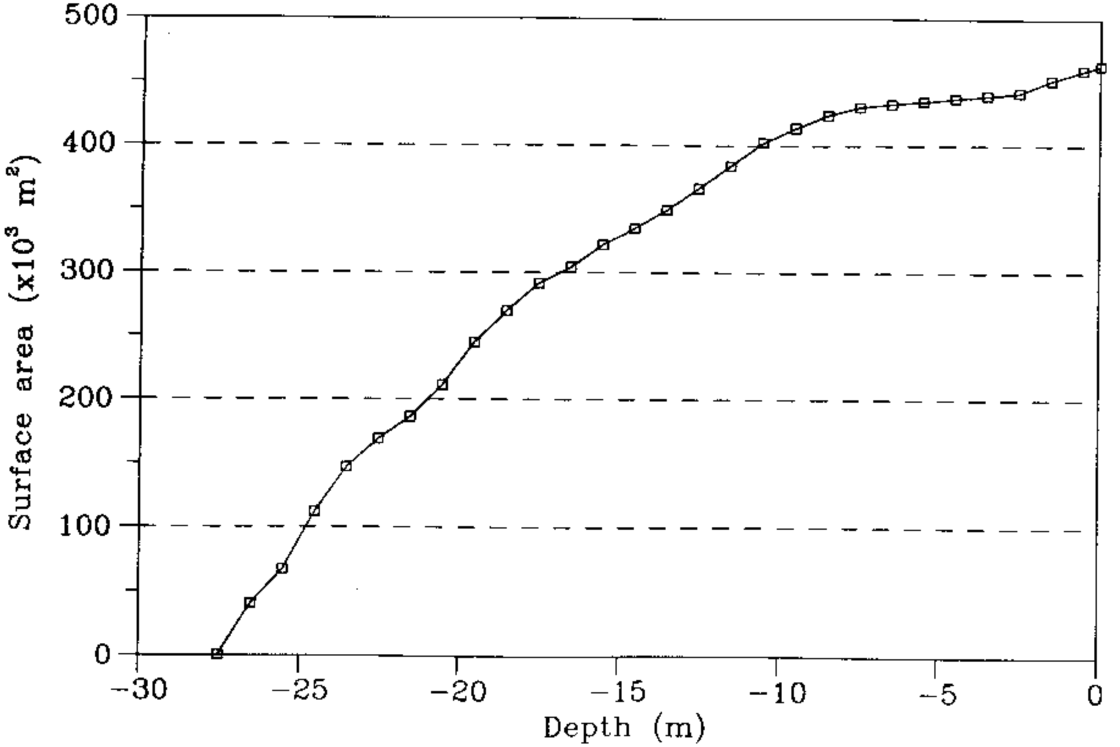


Figure 3.17 Variation of surface area of Nainital lake with depth.

Table 3.19 Surface area of Lake Nainital at different depths of water level.

Depth (m)	Area (m^2)	Depth (m)	Area (m^2)
0	463365	13.5	349038
0.5	459730	14.5	334423
1.5	451721	15.5	321538
2.5	441625	16.5	303846
3.5	439340	17.5	291280
4.5	437052	18.5	269615
5.5	434765	19.5	245000
6.5	432478	20.5	211346
7.5	430192	21.5	186376
8.5	423846	22.5	169230
9.5	413653	23.5	147115
10.5	402307	24.5	112307
11.5	383846	25.5	67307
12.5	365961	26.5	40769
		27.3	769
NOTE: 0m given above = 3.28 m in PWD Gauge on 22.11.1989			

Table 3.20 Sub-basin wise volume of Lake Nainital at different water levels

Depth (m)	Tallital sub-basin	Cumulative Volume (m^3)	Mallital sub-basin	Cumulative Volume (m^3)	Total lake Cumulative Volume (m^3)
0	-	-	-	-	8,581,714
0.5	_	_	_	_	8,336,862
1.5	_	_	_	_	7,880,420
2.5	_	_	_	_	7,434,634
3.5	_	_	_	_	6,994,371
4.5	_	_	_	_	6,556,175
5.5	_	_	_	_	6,120,267
6.5	_	_	_	_	5,686,646
7.5	_	_	_	_	5,255,311
8.5	_	_	_	_	4,828,296
9.5	_	_	_	_	4,409,556
10.5	_	_	_	_	4,001,588
11.5	_	_	_	_	3,608,547
12.5	_	_	_	_	3,233,679
13.5	_	_	_	_	2,876,212
14.5	_	_	_	_	2,534,507
15.5	_	_	_	_	2,206,547
16.5	_	_	_	_	1,893,896
17.5	_	_	_	_	1,596,355
18.5	106,831	418,484	150,940	897,493	1,315,977
19.5	87,438	311,653	141,314	746,553	1,058,206
20.5	69,578	224,215	130,586	605,239	829,454
21.5	58,463	154,637	121,764	474,653	629,289
22.5	48,761	96,174	112,643	352,889	449,063
23.5	35,042	47,413	95,952	240,246	287,659
24.5	12,371	12,371	74,961	144,294	156,665
25.5	-	-	53,487	69,333	69,333
26.5	Nil	-	15,713	846	15,846
27.3	Nil	-	133	133	133

NOTE: 0m given above = 3.28 m in PWD Gauge on 22.11.1989

3.4.2.1 Change in storage

Due to natural processes, the lake level changes with respect to time. However, the change in the water level in 24 hours (a day) or in a month or in a year indicates the change in storage of the lake for the selected time interval. The change in storage is an important parameter in water balance computations for the lake. The following equations have been used to calculate the lake surface area from the lake level data.

For LL between 3.81 and 2.44 m AREA = 6.86*LL + 437232
For LL between 2.44 and 1.80 m AREA = 14.35*LL + 418999
For LL between 1.80 and 1.49 m AREA = 6.12*LL + 433787
For LL less than 1.49 m upto 0 m AREA = 2.29*LL + 438698

where LL is the lake level in mm and AREA is the lake surface area in m^2. The results are then used to compute the volume between any two lake levels (falling within the range of 0 to 3.81m) by following the method given by Zumberge and Ayers (1964) described under section 3.4.2.

3.4.3 Thermal regime

The thermal regimes of lakes is basically a result of heat & momentum transfer across the surface of the lake and also due to the force of gravity acting on density differences with in the lake. In general the heat transfer at the surface tends to raise the temperature during summer and lower during winter. Therefore, the temperature difference at the top and bottom give rise to the mixing currents, but in summers these currents make a small depth cycle while during winter when the top layers become cooler in comparison to deeper, the sinking process starts which has greater mixing effects. Thus, the lakes of medium depth undergoes a complete mixing process and remains well mixed during winters while stratified during summers. This phenomenon is a consequence of several meteorological & other factors. The mixing and stratification behaviour of lakes is discussed below.

The depth versus temperature plots at different times of the year depict the seasonal changes in the temperature distribution in the lake. During winter, the lake may be considered as well mixed as there is negligible difference in the temperature of surface and the bottom waters. During spring, as a consequence of rise in air temperature thermal stratification is established. With the advent of summer and increase in solar radiation, the stratification is intensified. The mixing due to wind action, if any, is confined only to the surface layer (called epilimnion). The water of the bottom layer (called hypolimnion) remains cut off from the influence of wind. At this juncture the lake is said to be stably stratified. The layer that separates epilimnion and hypolimnion layers is called mesolimnion or metalimnion. This middle layer is also called thermocline. The term thermocline is also frequently used in the literature to denote the plane that passes through the points of maximum decrease in temperature with depth.

The stable stratification that persist during monsoon tends to become unstable during late monsoon and early winter, due mainly to falling air temperature resulting in net heat loss from the lake. During this period strong vertical mixing (convection) takes place. The process of convection, coupled with the increased action of wind results in the deepening of surface layer (sinking of thermocline) and complete mixing of lake takes place. This final phenomenon is called "winter overturn". Therefore, a moderately deep mountainous lakes undergoes two distinct stages namely complete mixed stage and a stratified stage of minimal vertical mixing.

The metabolism, physiology and behaviour of aquatic organism are related directly to the temperature of the aquatic environment. Extreme temperature restrict the growth and distribution of the plants, animals and microbes. Because of high specific heat of water, the temperature of large volumes of water change slowly. Therefore, large lakes tend to moderate the local climates. The temperature is one of the main causes of motion and mixing of lake's water therefore, the thermal structure, stratification of water masses and circulation patterns of lakes and reservoirs have profound effects on the nutrient cycling (spread of pollutants) and the distribution of dissolved gasses and the biota. For these and other reasons, the thermal structure and heat content of a lake must be known with some degree of accuracy in limnological studies.

The temperature of lake water at different depth were measured at centre points of southern and northern sub-basins of the lake viz. C2 and H2 as shown in Figure 3.2. The temperatures were measured using a thermometer (0.2°C least count), that is fixed inside the standard water sampler. Measurements were made at every 3m depth starting from the water surface.

The thermal characteristics of Lake Nainital are presented in Tables 3.21(a) and 3.21(b) and are shown in Figures 3.18 and 3.19. The surface water temperatures ranges from 9.2°C to 22.4°C to in the South basin, while in the North basin, it ranges from 8.4°C to 23.6°C. The temperature in the bottom waters of South basin ranges from 8°C to 9.4°C, and in North basin from 8°C to 9.2°C.The lake is thermally well mixed during the period from December to February. However the minor difference between the top and bottom waters are observed in December. During January the difference between epilimnion and hypolimnion water is 0.4°C only. The lake continues to loose energy and in February the minimum temperature condition is observed in the entire column of lake waters. The thermocline develops during march and the onset of stratification is seen to develop in the month of April. The stratification intensifies during May. The vertical temperature difference was maximum during June when the thermocline was at 7.5 m below surface. The stratification continues so upto November, when, the thermocline sinks to the bottom. And the thermocline disappears totally during the third week of December and the winter turn over takes place. It has been reported that algal blooms are associated with this process, as the nutrients from hypolimnion, hitherto unavailable to the biota in the surface waters, is released as a consequence of turnover. It can be seen from Figures 3.18 and 3.19, that the development of thermocline is first observed in March at 3m depth and later, it starts to sink and is observed at a depth of 13.5 m in November, before completely disappearing by the end of November.

3.4.3.1 Vertical exchange coefficient

The vertical exchange coefficient across the thermocline can be estimated by using the following equation (Reckhow & Chapra, 1983):

Figure 3.18 Temperature variation in Mallital sub-basin of Nainital lake during 1994-95

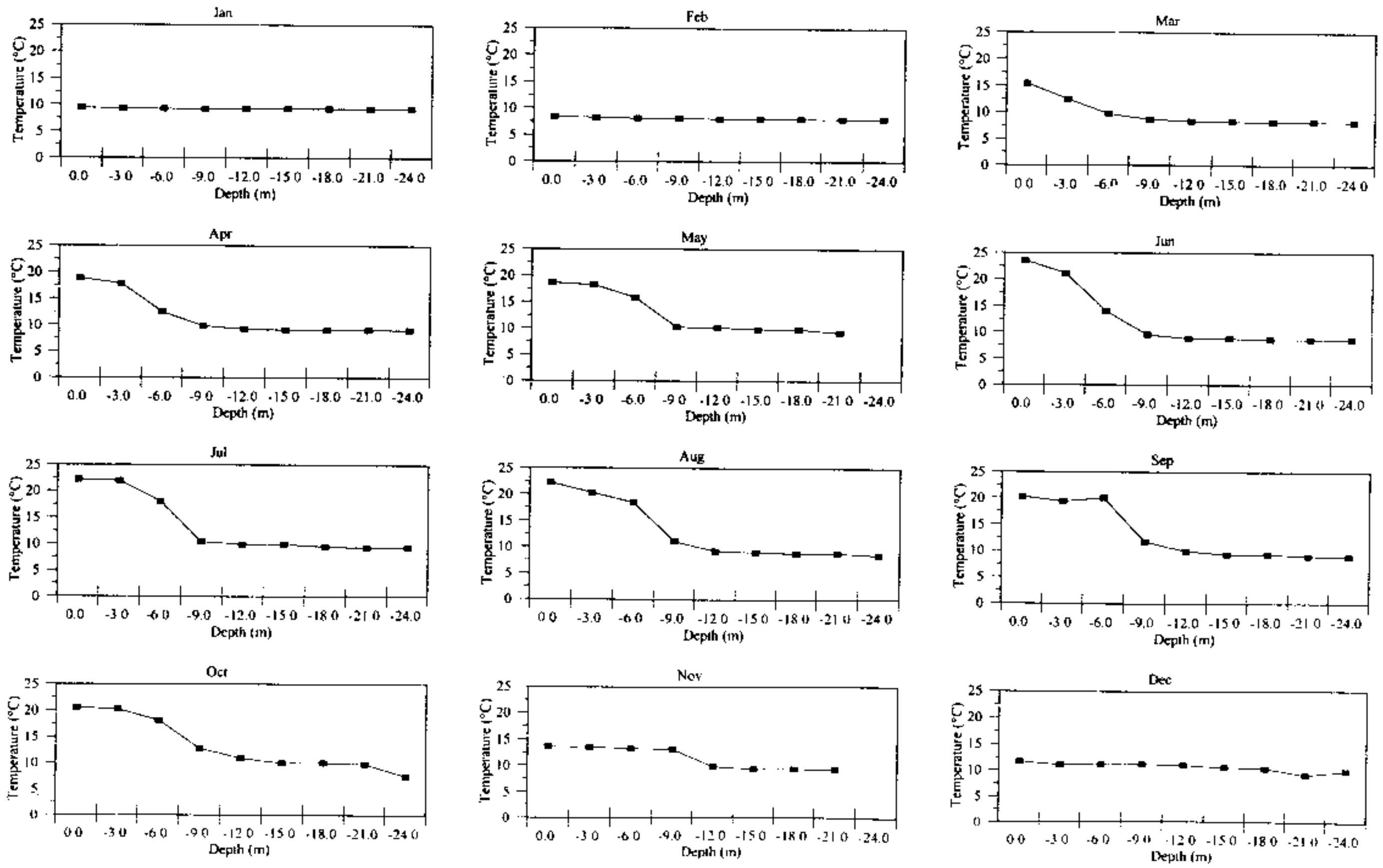
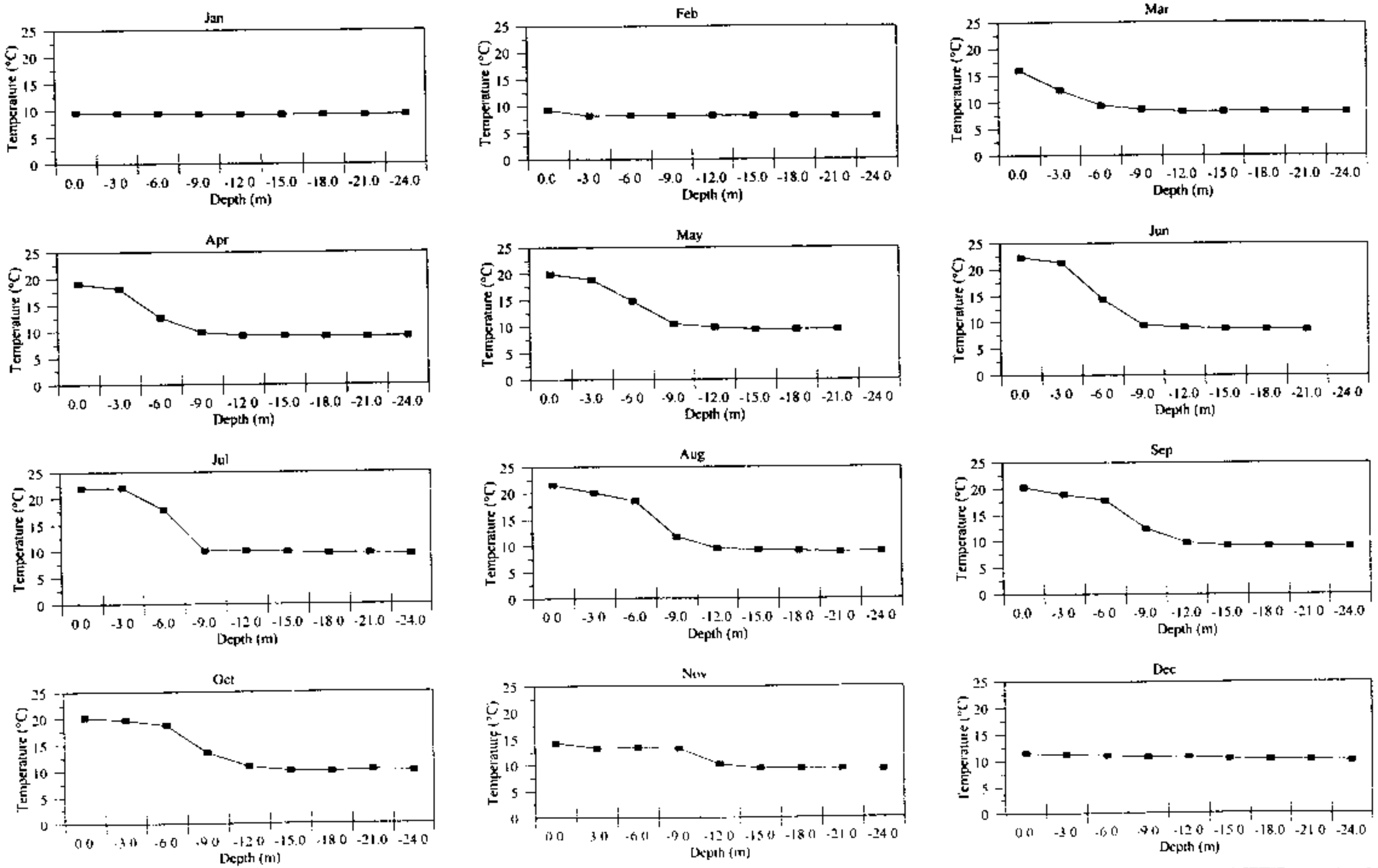


Figure 3.19 Temperature variation in Tallital sub-basin of Nainital lake during 1994-95



$$V_t = \frac{V_h}{2A_t\Delta_t}\,\frac{T_h^{n+1} - T_h^{n-1}}{T_e^n - T_h^n} \tag{3.15}$$

Where V_t = vertical exchange coefficient (m/d)

V_n = volume of hypolimnion (m^3)

A_t = area of thermocline (m^2)

Δ_t = time interval between temperature measurement (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)

The vertical diffusion coefficient (E_t) can be computed from the following equation (Reckhow & Chapra, 1983):

$$E_t = V_t\, Z_t\, \frac{10^4}{86400} \tag{3.16}$$

Where E_t = vertical diffusion coefficient (cm^2/sec)

V_t = vertical exchange coefficient (m/d)

Z_t = thickness of thermocline (m)

The coefficients thus evaluated are useful for gaining information on influence of groundwater on the hydrology of a lake by adopting stable isotope mass balance technique. The estimation of vertical exchange coefficient has been made using the methods decribed above and using the observed data, and are given in Table 3.22.

Table 3.22. Vertical exchange coefficients estimated from lake temperature data

Month	Vertical exchange Coefficient (m/d)	Remarks
January	-	well mixed
February	-	well mixed
March	0.017	Weakly Stratifed
April	0.007	Stratifed
May	0.005	Stratifed
June	0.004	Stratifed
July	0.003	Stratifed
August	0.007	Stratifed
September	0.008	Stratifed
October	0.005	Stratifed
November	0.031	Weakly stratified
Deecmber	-	well mixed

3.4.4 Surface Outflow

Surface outflow from the Nainital lake occurs only in the months of rainy season because, the water level in lake is maintained at Tallital bank by the PWD, Nainital with the aid of sluice gates. The lake outlet contains 5 gates, each 2.5 ft. wide and 1.25 ft. high, worked by screw type regulatory system. Each opening has an area of 3.75 square feet. All the sluices are opened simultaneously and to the same extent. The Hill Slide Committee of 1927 formulated some regulations regarding the opening of sluices and the same are still followed, the important features of which are given below.

If the gauge is below 9 ft., the sluices are not to be opened more than 6", and so long as the gauge is below 9.5 ft., the sluices as a rule are not to be opened more than 12", but in the event of abnormally heavy or continuous rainfall, the District Engineer/Surveyor must use his discretion in this matter. These regulations are mainly to avoid any damage to the geotechnically sensitive Balia ravine. The discharge through the lake outlet has been calculated by using the hydraulic equation for submerged rectangular openings. The equation for the gates installed in Lake Nainital has been given by Byrnes (1930) and the same has been adopted in the present study.

$$Q = 252780 \cdot (h_1 - h_2) \cdot \frac{\sqrt{h_1 + h_2}}{2} \tag{3.17}$$

where Q is the discharge in cubic feet per hour, h_1 and h_2 are the head to the bottom and top of the rectangular sluice opening in feet. The information on h_1 and h_2 have been obtained from the sluice gate operation records i.e., the time & extent of gate opening and the lake level at the time of sluice opening and closing. The records are well maintained by the UPPWD, Nainital. Further, under the present sluice operational procedures of the PWD, the sluices are never completely closed since the first day of its opening in any year until the 15th of October in that year, when the sluices are finally sealed. Leakage of lake is allowed during this period through a sluice opening of about half an inch. Therefore in the computation of the sluice discharges, the leakage has also been taken into consideration.

3.4.5 Chemistry of lake water

The water quality aspects of the Lake Nainital was monitored between February 1994 and July, 1996. Sampling was carried out during February, May, October and December, 1994; February, March, May, June, August, September, November and December, 1995; January, April and July, 1996. The number of samples collected from the lake varied from a minimum of 6 (June, 1995) and a maximum of 37 (March, 1995). However, keeping in view the need for control points, the centre points of North and South basins (C2 and H2) were sampled regularly during all the sampling periods. Sampling at any point in the lake was carried out, generally, at three different depths viz., the surface, middle and bottom of the lake. During some sampling periods, sampling was carried out at every three meter depth interval, starting from surface to the bottom.

Apart from the analyses of lake waters - as noted above - other water sources in the surrounding region were also studied. The hydrochemistry of the groundwater is described separately (Section 3.5.3). The number of samples collected from the lake for analyses during the different sampling campaigns is given in Table 3.23.

Table 3.23 Details of samples collected from Lake Nainital

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		

In the year 1994, sampling was carried out during 4 months, which was considered to be representative of the four different seasons. However, realising the dynamics and the possible short hydraulic residence time of the lake water, the sampling was carried out during 8 months (as shown in the table) in the year 1995. In the last year of the investigations i.e. 1996, sampling was carried out in January, April and July with a view to complement the available information on the lake's quality.

3.4.5.1 Materials and methods

The samples from the lake were collected using a depth sampler (Hydrobios make) of 2 L capacity with a built in thermometer of 0.2°C least count. The electrical conductivity (E.C.)

and pH were measured in-situ using a 5-parameter field water testing kit. In the early stages the dissolved oxygen was also measured in the field. But later comparison with laboratory data showed the non-reliability of the field DO values. Hence, the in-situ measurements were restricted only to temperature, pH and E.C.

As stated above samples were collected from fixed points (Figure 3.2). To understand the hydrodynamics of the lake and also the spatial variation of the water characteristics within the lake, samples were collected at every three meter depth, starting from the surface upto the bottom. The collected samples were mainly utilised for three different aspects, viz. a) In-situ measurement of physico-chemical characteristics, b) Isotope composition and c) water quality aspects. While in-situ measurements and samples for isotope characteristics were made every three meters, samples for water quality of aspects were made only at three or four different depths viz. Epilimnion (0.5 m below surface), Mesolimnion (6 - 9 m below surface, depending on the season) and hypolimnion (1 to 4 m above the bottom, depending on the season).

The samples collected for the water quality aspects were first filtered using an 0.45 μm Whatman filter and preserved. Each sample was filled in three different bottles. The first one (1 L) was unpreserved for alkalinity, chloride, sulphate and Orthophosphate - phosphorus determination. The second one (1 L) was preserved with concentrated sulphuric acid for total hardness, and calcium hardness determination. The third one (60 mL) was preserved with concentrated nitric acid for Sodium and Potassium. The samples were then transported to the National Institute of Hydrology laboratory at Roorkee, and were immediately analysed giving priority to those parameters that are to be determined using the unpreserved samples. Other parameters were analysed within a week time. Apart from the above three bottles, one sample was collected from each sampling point for Dissolved Oxygen (DO) determination. The DO bottle of 300 mL capacity was filled upto the brim and preservatives were added according to Winkler's method.

The samples for stable isotope analyses were collected at every three meters depth in 20 mL plastic vials and tightly sealed with tapes and were transported to Bhabha Atomic Research

Centre (BARC), Mumbai. Samples were also collected for environmental tritium analyses and sent to BARC, during the month of December, 1994. The physico-chemical parameters that were analysed in the water samples include: a) Physical parameters viz., Temperature, pH and Electrical Conductance b) Major ions viz., Ca, Mg, Na, K, HCO_3, SO_4 and Cl, c) Nutrients such as PO_4-P and NO_3-N, d) Stable isotopes Oxygen-18 and Deuterium, and e) Environmental tritium. The methods followed at the NIH water quality laboratory for analyses of different parameters discussed above are as follows:

Chloride was estimated by Argentometric method in the form of silver chloride. Alkalinity was determined by Titrimetric method using phenolphthalein and methyl orange indicators. Total hardness and calcium hardness were determined by EDTA Titrimetric method while magnesium hardness is calculated by deducting calcium hardness from total hardness. Calcium (as Ca) may be calculated by multiplying calcium hardness with 0.401 while magnesium (as Mg) by multiplying magnesium hardness with 0.243. Sodium and potassium were determined by flame-emission method using Flame Photometer. Nitrogen in the form of nitrate was determined by using Spectrometric methods. 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. Trace elements were determined by atomic absorption or atomic emission methods. All the above analyses for water chemistry were carried out as per appropriate methods prescribed in the APHA Handbook (1985).

3.4.5.2 Analyses of data and accuracy

It is established that the sampling techniques, preservation methods and length of storage time before analysis have certain effect on the accuracy of the results. Under optimum conditions the results for major constituents of the water have an accuracy of ± 2% to ±10%. Solutes present in concentrations above 100 mg/L generally can be determined with an accuracy of better than ± 5%. For solutes present in concentration below 1 mg/L, the accuracy is not better than ± 10% and can be poorer. One way of determining the accuracy of major dissolved constituent values in a reasonably complete chemical analysis of a water sample can be accomplished by checking anion - cation balance (electro-neutrality). Another procedure for checking the accuracy is possible using specific conductivity and dissolved solids determinations. The total of milli-equivalents per litre of either anions or cations multiplied by 100 usually agrees approximately with the electrical conductivity in μmhos/cm (Hem, 1985). Considering the results of the repeat analysis and duplicate analysis of samples and electro-neutrality for estimating analytical accuracy, it is found that the overall accuracy of the analyses is within ± 5% error limit.

3.4.5.3 Characterisation

As described in section 3.4.3 the lake, based on its thermal characteristics, can broadly be divided into epilimnion and hypolimnion. However, this division is non-existent during the winter when the whole lake is well mixed. Although the epilimnion and mesolimnion have different concentrations of certain chemical ions during the stratified period the lake, in general, is of Magnesium Bicarbonate type. Based on the thermal characteristics of the lake, prevailing weather conditions and hydrological characteristics of the lake four different seasons may be envisaged (Table 3.24) for studying the lake water chemistry.

Table 3.24 Thermal characteristics of Lake Nainital and Hydometeorologic characteristics of the lake basin during different seasons

Season	Thermal stratification	Δe-h (°C)	Precipitation	Dominant hydrologic process
Winter	Well-mixed	< 2°	Occasional snowfall	Snowmelt
Pre-monsoon	Strongly stratified	8 - 11°	Low Rainfall	Evaporation
Monsoon	Well stratified	11 - 15°	High rainfall	Surface inflow and outflow
Post-monsoon	Weakly stratified	5 - 10°	Negligible rainfall	Sub-surface inflow

The observed range and mean concentrations of different anions/cations along with the corresponding statistical errors in the estimated mean of the concentrations in the samples collected from the epilimnion 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 Tables 3.25a to 3.25d.

Table 3.25a Average values of Electrical Conductance, pH and Total Hardness during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital

Parameter	units	Season	Number of samples	Range		Mean	σ
				Min	Max		
EC	μS/cm	winter	71	370	680	500	50
		Pre_E	21	440	890	600	130
		Pre_H	60	360	860	620	140
		Mon_E	16	380	650	460	90
		Mon_H	35	480	850	600	110
		Post_E	14	510	590	550	30
		Post_H	30	580	770	660	30
pH	-	winter	51	6.4	8.7	7.5	0.7
		Pre_E	14	7.8	9.4	8.5	0.6
		Pre_H	45	7.2	8.9	8	0.4
		Mon_E	15	6.9	9.6	8.6	0.8
		Mon_H	30	7.0	9.4	7.7	0.6
		Post_E	14	7.5	8.8	8	0.3
		Post_H	30	6.7	8.3	7.2	0.3
Hardness	mg/l	winter	72	300.0	422.0	358.5	34.3
		Pre_E	21	280.0	373.0	329.9	29.7
		Pre_H	61	324.0	391.0	344.4	12.7
		Mon_E	16	240.0	348.0	264.1	25.4
		Mon_H	35	236.0	394.0	339.1	25.9
		Post_E	14	244.0	416.0	320.4	65.5
		Post_H	30	274.0	442.0	351.7	50.8

Table 3.25b Average values of Alkalinity, Sulphate and Chloride during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital

Parameters	units	Season	Number of samples	Range		Mean	σ
				Min	Max		
Alkalinity	mg/l	winter	52	230.0	350.0	275.4	23.2
		Pre_E	18	197.0	268.0	237.3	22
		Pre_H	55	234.0	304.0	259.5	12.2
		Mon_E	16	170.0	208.0	195.6	9.9
		Mon_H	35	242.0	318.0	268.4	17.4
		Post_E	14	174.0	235.0	210.1	22.3
		Post_H	30	218.0	274.0	256	12.9
SO_4^{2-}	mg/l	winter	56	49.0	145.0	93.3	15.4
		Pre_E	21	68.0	126.0	89.6	17.1
		Pre_H	61	64.0	135.0	90.7	16.6
		Mon_E	14	55.4	120.0	89.1	14.6
		Mon_H	29	56.4	120.0	94.1	11.7
		Post_E	14	74.1	117.7	89.5	10.5
		Post_H	25	45.3	98.0	80.2	9.5
Cl^-	mg/l	winter	66	3.0	32.0	14.4	7.7
		Pre_E	17	10.0	26.0	15.7	3.5
		Pre_H	55	6.0	22.0	14.7	3
		Mon_E	16	0.0	28.0	14.1	9
		Mon_H	34	0.0	72.0	20.5	14.5
		Post_E	14	6.0	37.0	15.5	7.3
		Post_H	30	5.0	21.0	13.1	3.4

Table 3.25c Average values of Phosphorous, Nitrogen and Sodium during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital

Parameters	units	Season	Number of samples	Range		Mean	σ
				Min	Max		
P(PO-4)	mg/l	winter	70	0.11	1.47	0.29	0.23
		Pre_E	13	0.02	0.16	0.11	0.04
		Pre_H	45	0.08	0.78	0.26	0.13
		Mon_E	14	0.00	0.36	0.15	0.09
		Mon_H	29	0.00	0.69	0.32	0.16
		Post_E	14	0.00	0.59	0.1	0.2
		Post_H	30	0.10	0.60	0.29	0.14
N(NO-3)	mg/l	winter	67	0.0	5.0	1.05	0.98
		Pre_E	8	0.0	0.0	0	0
		Pre_H	33	0.0	1.3	0.8	0.47
		Mon_E	12	0.0	0.0	0	0
		Mon_H	25	0.0	3.3	1.49	0.99
		Post_E	14	0.1	1.1	0.46	0.22
		Post_H	30	0.0	7.0	2.67	1.42
Na^+	mg/l	winter	67	5.4	14.4	9.8	2.4
		Pre_E	14	9.4	13.6	11.4	1.3
		Pre_H	45	8.0	14.6	11.5	1.5
		Mon_E	14	10.8	19.0	15.1	1.9
		Mon_H	29	9.0	21.0	16	3.4
		Post_E	6	13.8	39.8	23.7	11.5
		Post_H	10	16.9	39.8	20.6	6.8

Table 3.25d Average values of Potassium, Calcium and Magnesium during different seasons observed in the epilimnion and hypolimnion zones of Lake Nainital

Parameters	units	Season	Number of samples	Range		Mean	σ
				Min	Max		
K^+	mg/l	winter	68	3.1	6.7	4.4	0.6
		Pre_E	14	4.1	9.0	4.8	1.2
		Pre_H	45	4.0	5.2	4.5	0.3
		Mon_E	14	3.4	7.8	5.5	1.3
		Mon_H	29	3.6	8.1	6	1.3
		Post_E	6	3.5	4.9	4.1	0.6
		Post_H	10	4.1	5.0	4.7	0.2
Ca^{2+}	mg/l	winter	71	28.1	73.8	53.4	8.4
		Pre_E	21	32.9	56.9	42.7	6.5
		Pre_H	61	36.1	80.2	49.3	7.6
		Mon_E	16	21.6	45.7	33.3	6.4
		Mon_H	35	26.5	64.2	39.1	10.7
		Post_E	14	26.5	57.7	42.6	12.5
		Post_H	30	42.5	80.2	59.9	11.5
Mg^{2+}	mg/l	winter	71	32.0	96.0	54.6	8.2
		Pre_E	21	44.7	63.2	54.5	5.5
		Pre_H	61	41.3	61.2	53.7	4.1
		Mon_E	16	31.6	53.2	42.5	5.5
		Mon_H	35	45.7	73.9	59.5	6.9
		Post_E	14	40.8	70.0	52.1	9
		Post_H	30	38.4	69.0	49.2	7.2

From the above table it is seen that except for Ca, Mg and HCO_3, all the other chemical parameters do not show any significant variation in their mean concentrations during different seasons.

3.4.5.4 Stable isotope ratios

Samples for stable isotope analyses were collected from the lake during various sampling campaigns as detailed in Section 3.4.4. The samples were collected in 20ml plastic vials and were tightly sealed to avoid evaporation losses and were transported to the laboratories. Most of the lake samples were analysed for the stable isotope ratios using the techniques outlined in Section 3.2.6.1 at the IRMS facility at the Hydrology and Tracer Section, Isotope Division of Bhabha Atomic Research Centre, Mumbai. However, one set samples collected during September, 1996 was analysed at Environmental Isotope Laboratory at University of Waterloo, Ontario, Canada. The brief results (arranged chronologically) are given in Table 3.26.

Table 3.26 Mean observed $\delta^{18}O$ and δD in the epilimnion and hypolimnion zones of Lake Nainital during different sampling campaigns

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 error values (sigma) presented adjacent to the mean values in the above table represent the variation in the isotopic delta values at different points in the lake. The values given with in the parentheses are the number of samples used for the estimation of the mean.

The equation for the best-fit line using the measured values of oxygen and hydrogen stable isotope contents of lake water samples collected during different seasons is as follows:

$$\delta D = (7.1 \pm 0.4) \cdot \delta^{18}O + (2.3 \pm 2.6) \quad (n = 131;\ r = 0.74) \tag{3.18}$$

The above equation is very close to that of the LMWL (section 3.2.6.2). This indicates that the lake probably does not undergo any significant evaporative enrichment. The isotope data at best could be used to qualitatively describe the system and may not be very much useful in quantitative estimations of the lake water balance components.

3.4.6 Lake evaporation

The estimation of evaporation is rather complicated. Several equations for estimating evaporation from free water surfaces exist in literature. In a recent and most comprehensive article, Winter *et al* (1995) have evaluated 11 equations for the estimation of lake evaporation. Of the three types of approaches to the estimation of evaporation, viz. energy balance, mass transfer and combination (Penman type), the first one yields more reliable results, if data on all associated energy terms are available. Alternatively, if sufficient data is available, the combination methods yield better results. Based on the experience in Lake Nainital, the Penman method (Jensen, 1974) was chosen. The equation for the estimation of evaporation (E) is:

$$E = \frac{\Delta}{\gamma + \Delta} \cdot (Q_N + G) + \frac{\gamma}{\gamma + \Delta} \cdot 15.36 \cdot (0.5 + 0.01 \cdot U_2) \cdot (e_0 - e_a) \tag{3.19}$$

Where, γ and Δ are the weighting factors. The following equation can be used for the computation of Δ, that is the slope of the vapour pressure versus temperature curve and is expressed in mb/°C.

$$\Delta = \frac{25083}{(T + 237.3)^2} \cdot \exp\left(\frac{17.3 \cdot T}{T + 237.3}\right) \qquad (3.20)$$

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

$$\gamma = \frac{0.61 \cdot P}{1000} \qquad (3.21)$$

where P is the mean atmospheric pressure in mb. The atmospheric pressure data has been collected from UP State Observatory at Manora 2 km south of the lake site. Q_N is the net solar radiation. Since the solar radiation data for the site was not available, the net radiation was computed using the following relationship:

$$Q_N = Q_S \cdot (1 - \alpha) - Q_{LN} \qquad (3.22)$$

where, α is the albedo (reflection coefficient), Q_S is the global solar radiation and Q_{LN} is the net long wave radiation. The α for the present study was computed by using the values presented for different amount of cloud cover (Ter-Makaryantz, 1960). The sunshine hours data collected by Uttar Pradesh State Observatory (UPSO) located about 2 km south of the lake was used to generate the cloud cover data. Rao *et al* (1971), after analysing meterological data collected from IMD Stations from all over India, proposed the following equation for the computation of Q_S.

$$Q_S = Q_A \cdot [0.325 \cdot COS\Phi + 0.385 \cdot (\frac{n}{N})] \qquad (3.23)$$

where Φ is the latitude and n and N are actual and maximum possible sunshine hours, respectively. Q_A, the extraterrestrial radiation and N for the site were computed using the Duffie-Beckman equations (Allen, 1996). Q_{LN} was computed by the following equation (Shuttleworth, 1992):

$$Q_{LN} = - f\ e'\ \sigma\ T^4 \qquad (3.24)$$

where, σ is the Stefan-Boltzmann constant, T is the absolute temperature in Kelvin, f is the adjustment factor for cloud cover, which is roughly equal to (0.1+0.9*n/N) (Singh, 1992) and ϵ' is the emissivity factor, which was computed using the Idso-Jackson equation. (Nokes, 1995). U_2 is the wind velocity in m/s at 2m above surface. e_0 and e_a are the 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 (Linsely et al., 1975):

$$e_0 = 33.8639 \cdot [(0.00738\, T_0 + 0.8072) - 0.000019 \cdot |1.8\, T_0 + 48| + 0.001316] \quad (3.25)$$

The actual vapour pressure e_a can be computed by multiplying the relative humidity with the saturated vapour pressure at air temperature. The parameters, required for the Penman equation, viz., air & water temperatures and relative humidity were measured at the lake site. Data was also collected from a USWB Class-A Pan installed at the lake site. The estimated & measured parameters used for the computation of evaporation and the values of estimated evaporation from lake Nainital are presented in Table 3.27.

Table 3.27 Values of measured and estimated parameters used in the estimation of evaporation from Lake Nainital by Penman Method for the year 1995.

1995	Atm. Press.	Sun Shine	Air Temp.	Lake Temp.	Wind Velo.	Rel. Humi.	Alpha Coeff.	Net Radn.	Lake Evap.
Units	mb	Hrs.	°C	°C	m/s	%	-	ly/day	mm/d
Jan	812.78	6.76	4.43	8.82	3.19	58.42	0.087	145.19	2.14
Feb	812.37	6.22	6.41	8.78	3.23	53.42	0.074	191.07	2.68
Mar	813.48	8.61	10.05	12.08	4.89	47.96	0.068	278.47	3.80
Apr	812.82	6.47	15.31	15.96	4.15	47.30	0.058	310.60	4.38
May	812.26	8.25	20.22	19.90	4.73	61.95	0.061	368.75	5.15
Jun	809.57	2.41	21.78	22.47	3.60	68.88	0.056	288.31	4.12
Jul	809.27	3.10	18.77	21.51	2.75	85.74	0.057	295.06	3.90
Aug	810.02	2.91	18.41	20.93	2.72	87.43	0.056	272.73	3.58
Sep	811.97	4.53	17.18	20.13	2.25	81.10	0.064	256.56	3.40
Oct	814.50	8.28	15.32	18.42	2.12	72.64	0.062	239.31	3.22
Nov	813.86	8.94	12.22	14.64	1.75	51.17	0.076	169.25	2.60
Dec	814.15	7.76	8.67	11.17	1.82	53.81	0.087	129.19	2.08

3.5 Hydrogeology

The systematic hydrogeological studies in Nainital basin have not been conducted earlier, mainly due to the abundant surface water availability. However, the intensive geological surveys conducted to identify the causes of recurring landslides and to delineate hazardous zones by earlier investigators do give some insight to the probable hydrogeologic setting of the basin. As noted in Section 3.1 on the Geology of the Lake basin, most part of the basin is characterised by calcareous and pyritiferous shales. Most of the lake catchment area lying north of the Lake Fault is characterised by Manora shales that dip towards the lake. The average dip of the shales / slates in the Sher-ka-Danda is 40° WSW, and in the Alma Hill 40-48° SW. Holland (1896) noted that the main cause of the slope instability is the rain water seeping through the bedding planes thereby weakening the slope strength. Similar observation was made by the same investigator for the dolomitic sandstone near the East Laggan to the west of the Lake.

The fact that the rain water infiltrating the soil cover in the Nainital basin, moves preferentially along the bedding planes is corroborated by the absence of springs in the Rosamond's ravine, from where the infiltrated water finds its way to the Glenmore ravine marked by springs (Holland, 1896). It is possible that the infiltrating waters do not percolate to greater depths and that they either discharge into the lake or as issue out as seasonal springs. During the investigations carried out as part of the present study also, several seasonal springs were observed, the flow of which diminishes with time before completely drying-up by the middle of post-monsoon season.

To understand the groundwater conditions in the basin that is characterised by multitudes of faults and fractures, it is imperative to have detailed geological sections. Geological Survey of India has drilled about 7 boreholes during 1967-68 & 1976-78 as part of the investigations carried out to the study the hill slope instability. Out of these seven boreholes, four were constructed with perforated steel pipes to facilitate monitoring of the groundwater level. Unfortunately, all the wells except one (DH#2), were unavailable for any further investigation. However, the lithologs reported by the investigators (Hukku et al., 1967; Ashraf, 1978) are useful in gaining valuable information on the thickness and nature of different rock strata. The ENE-WSW cross-section of the rock formations between St. Loe and the Copacabana Restaurant near the lake is shown in Figure 3.20. The section has been prepared using the information available in the Hill slope instability investigations reports of Ashraf (1978) and Holland (1896). It is seen from the figure that the shales dip towards the lake at an angle of ~40° and the groundwater level follows the surface topography. The top 5-6m is characterised by overburdens (talus and slide debris). The next 12-35m is characterised by weathered shales of Lower Krol. Below this depth the rocks are reportedly firm in nature. Between 29 and 34 m depth in Drill hole No. 3, a shear zone has been reported by Ashraf (1978). The depth of the overburden increases towards the WSW part of the section, i.e. near the foot hills. The depth of debris at Drill hole No. 5 is about 35m from ground level.

In striking contrast to the setting of rock formations in the Sher-ka Danda hill described above, the Flats or landslide deposit lying immediately to the north of the lake is characterised by landslide debris, nala deposits and lake deposits upto a depth of 116m below ground level. Below the depth of 116m, shales / slates of Middle Krol have been reported by Ashraf (1978) that confirms the presence of Lake Fault. On these Flats the three tube wells of the UP Jal Nigam/ Jal Sansthan are located and the tube wells are used for pumping groundwater in large quantities for the town water supply (cf Section 3.5.3).

Further north/northwest of the lake and Flats the Lower Krol shales continue, the western boundary of the outcrop marked by the Nainadevi Drain, considerable length of which follows the Lake Fault. To the south of the Lake Fault also some area is marked by the Manora Shales. The rest of the west and southwest part of the lake basin is characterised by limestones and dolomites that generally dip towards N / NW at varying angles (20°-60°). The exception to this is the south western part of the lake basin where the beds dip towards south. This has interpreted as a synclinal fold (Middlemiss, 1890; Valdiya, 1988). The dolomite occurring in the lake catchment is marked by swallowholes, such as the one present north of the Sherwood College. However, this massive dolomite and associated limestones have not developed all the characteristic features of a karstic terrain, such as vertical shafts (or natural wells). There are a few cave like structures observed in and around the lake basin. One such cave the near the

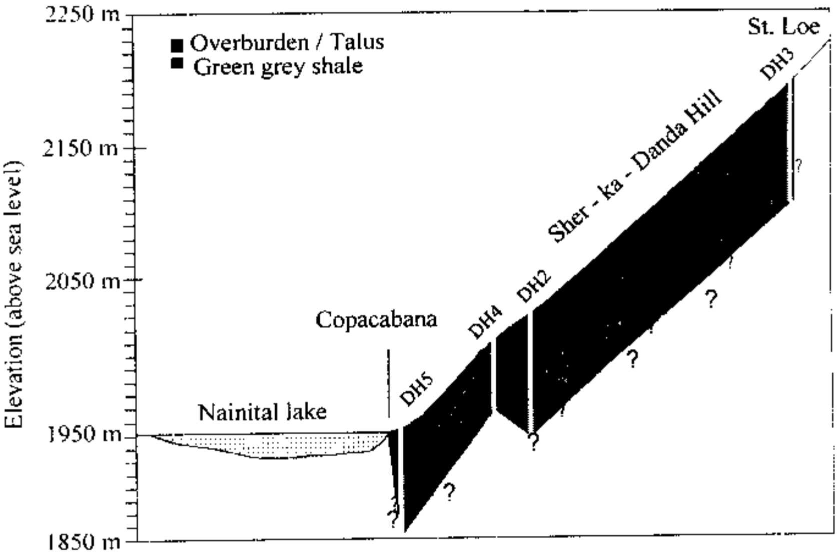


Figure 3.20 Geological cross-section along Copacabana and St. Loe in the Sher - ka - Danda hill, Nainital. Data source: Ashraf, 1978

Pashandevi temple (located on the western bank of the Lake) is cited by A. K. Sharma (Kumaun University, Personal communication) as a resultant of enlargement of joints by solution activity. Other cave like structures are seen in the western slopes of Balia ravine (Gupha Mahadev Temple) and in the northern slope of Sukhatal. Although, the latter one is fairly large in size, ~10m long and about a meter wide with smoother inner wall marked with smaller openings extending laterally, presently it appears dry with no apparent flow of water. Another striking feature of the dolomitic - limestone terrain in the lake basin, is that it is completely devoid of any spring in contrast to the Manora shales in the northern and eastern parts of the Lake basin.

The Nainital basin is a structurally and stratigraphically complex area. In an area such as this the stratigraphic positions of the carbonate rocks in relation to the less permeable overlying rocks or non-permeable underlying rocks. and also the geologic structures that control the recharge and discharge areas influences the gross behaviour of the aquifer system. Among the sedimentary rocks, shales are the least productive and carbonate rocks vary widely in their water yield characteristics. The variations are mainly due to the extent which the carbonate rocks have been subjected to the solvent action of the circulating waters (Stringfield and Le Grand, 1969). The older carbonate rocks are generally compact and impervious except along bedding planes and joints where solution by circulating water may increase the permeability. However, if the limestone is folded into anticlines or other structures water may move approximately parallel to the dip of the underlying and overlying confining beds. Under these conditions water may move down dip from the recharge area on the crest of an anticline and up dip in the flank of an adjacent syncline to a discharge area. The joints and to certain extent the bedding planes may have a pronounced effect on the solution patterns and movement of water in carbonate rocks. Faults may affect the lateral movement of water and the solution of carbonate rocks, especially if the water bearing beds are faulted against relatively impervious rocks. In such a case the springs occurring along the fault zones may be the principal natural discharge of the productive aquifer. (Stringfield and Le Grand, 1969). The groundwater may also move along the solution channels along the fault zones. However, Davies (1960) opined that the groundwater movement in carbonate rocks are controlled by joints and the faults exert very little influence unless the faults and the regional hydraulic gradient are in general alignment.

3.5.1 Aquifer properties

Aquifer tests such as pump tests yield reliable information on the aquifer properties. However, due to the absence of essential conditions viz. existence of a pumping well and an observation well in the near-by area, pumping tests could not be conducted. However, an alternative method has been attempted for obtaining the required information, by conducting a slug test. Karanth (1987) has observed that the slug tests are useful to estimate the aquifer parameters only when no facility exists for carrying out pump tests and that the they only provide an preliminary information on the aquifer characteristics. The slug test is an alternative to the aquifer test and is performed in a small diameter monitoring well (Fetter, 1994). There are two different methods for conducting the slug tests, viz. Cooper - Bredehoeft - Papadopulus Method and Hvorslev method.

In the present study, Hvorslev method has been adopted, since the first method is applicable for confined aquifers and latter for unconfined conditions. The method involves adding a known quantity of water into the well casing or a by rapidly lowering a piece of metal, called a slug, into the well. The water level will rise as a consequence of the introduction of slug and after certain time the water level will gradually fall back towards the static water level. The water levels in the well are measured prior to and immediately after the introduction of the slug, and also at timed intervals till the water level falls back to 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 some time, t, after the slug is lowered is h. If the ratio h/h_0 at different time t are plotted against the corresponding time (t) on a semilogrithmic paper, the time - drawdown data falls on a straight line. If the length of the piezometer is more than 8 times the radius of the well screen then (Fetter, 1994):

$$K = \frac{r^2 \ln (L_e/R)}{2 L_e T_0} \qquad (3.26)$$

where K = hydraulic conductivity (m/day)

r = radius of the well casing (m)

R = radius of the well screen (m)

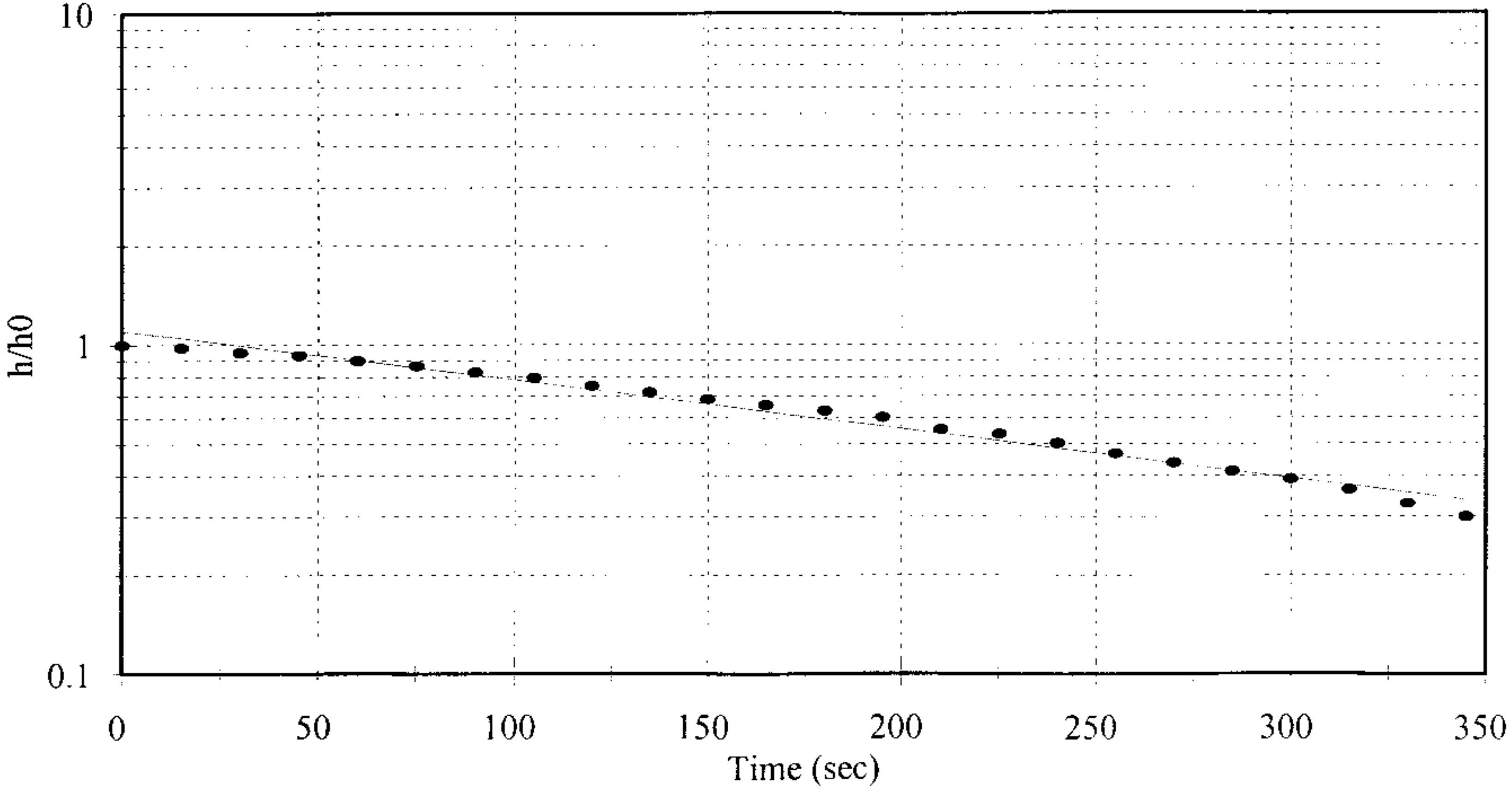
L_e = length of the well screen (m) and

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

In the present study the piezometer drilled by GSI investigators (DH#2) in the Sher-ka-Danda slope near the Spring Cottage (above the spring) was used to conduct the slug test. The total length of the well (constructed with perforated pipes for the complete length) is 91.72m and the radius is 0.019m. The static water level in the well prior to the test was 17.00 m below measuring point. 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 (Hydrobios™ make) that produces a audible signal at the contact point and the accuracy of the equipment is ±1.0mm. The water level rose to a height of 0.29m immediately upon introduction of the slug. The total time taken for the water level to fall back to the static water level was ~9.75min., and the T_0 calculated from the h/h_0 - t semi-log plot (Figure 3.21) is 311s or 0.003597 days. The hydraulic conductivity estimated from the above data using equation (3.26) is $4.64 * 10^{-3}$ m/day or $5.4 * 10^{-8}$ m/s. Assuming an aquifer thickness of 100m the aquifer transmissivity may be estimated as $T=Kb$, where b is the aquifer thickness. The estimated T is about 0.464 m^2/day. The estimated K and T values for the Manora shales are less atleast by one magnitude, compared to those of other shales formations occurring in other parts of India (the mean K and T values are 0.04 m/d and 5.5 m^2/d, respectively for the Semri shales in Andhra Pradesh; Karanth, 1987). However, the estimated K values are slightly higher than the range of values quoted for shales by others ($2 * 10^{-9}$ m/s to $1 * 10^{-13}$ m/s; Domenico and Schwartz, 1998).

The storage coefficient of the aquifer may be estimated from the well hydrograph analyses. Generally a well hydrograph follows a trend similar to that of a stream hydrograph, with a rising limb corresponding to period of recharge, followed by a recession limb corresponding to period of discharge from groundwater storage. The recession limb is characterised by an initial steep slope representing rapid drainage from storage, followed by a gentle limb tending to stabilise after a prolonged period of drainage. The physical process of

Figure 3.21 Plot of h/h0 versus time for estimation of K in Manora Shales by slug-test using Hvorslev method



releasing water from storage in the aquifer may be defined by the following equation (Karanth, 1987):

$$(h - h_m) = (h_0 - h_m)\ e^{-\alpha t} \qquad (3.27)$$

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

h_0 is the water level at the start of the recession,

h_m is the water level when the rate of recession is nil (at the end of post monsoon period),

α is the recession coefficient that may be approximated to S (storage coefficient).

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

$$h = h_0\ e^{-\alpha t} \qquad (3.28)$$

Rearranging the above equation we get,

$$\alpha = \frac{\log\ (h_0)\ -\ \log\ (h)}{t} \qquad (3.29)$$

from which the value of α and therefore, S of the aquifer may be estimated. The groundwate level data pertaining to the period 1990-91 for the wells DO#2 and DO#4 were obtained from UPPWD and the same were analysed using Equation 3.29. The estimated S ($\approx\alpha$) are $1.56 * 10^{-4}$ from the water level data from well DH#2 (Figure 3.22) and $1.34 * 10^{-4}$ from that of DO#4 (Figure 3.23). Since the estimated values are very close to each other, a mean value of $1.45 * 10^{-4}$ may considered as a true representation the aquifer storage coefficient. In case of water table or unconfined aquifer, the concept of storage coefficient is considered analogous to specific yield (Karanth, 1987). Expressed in percentage the specific yield of Manora shales is ~0.015%, that is much lower compared to other shale formations (Cuddapah shales, Andhra Pradesh have a mean specific yield of 3%; Karanth, 1987).

Figure 3.22 Well hydrograph of DH#2 located in Manora Shales below the spring near Spring Cottage, Nainital.

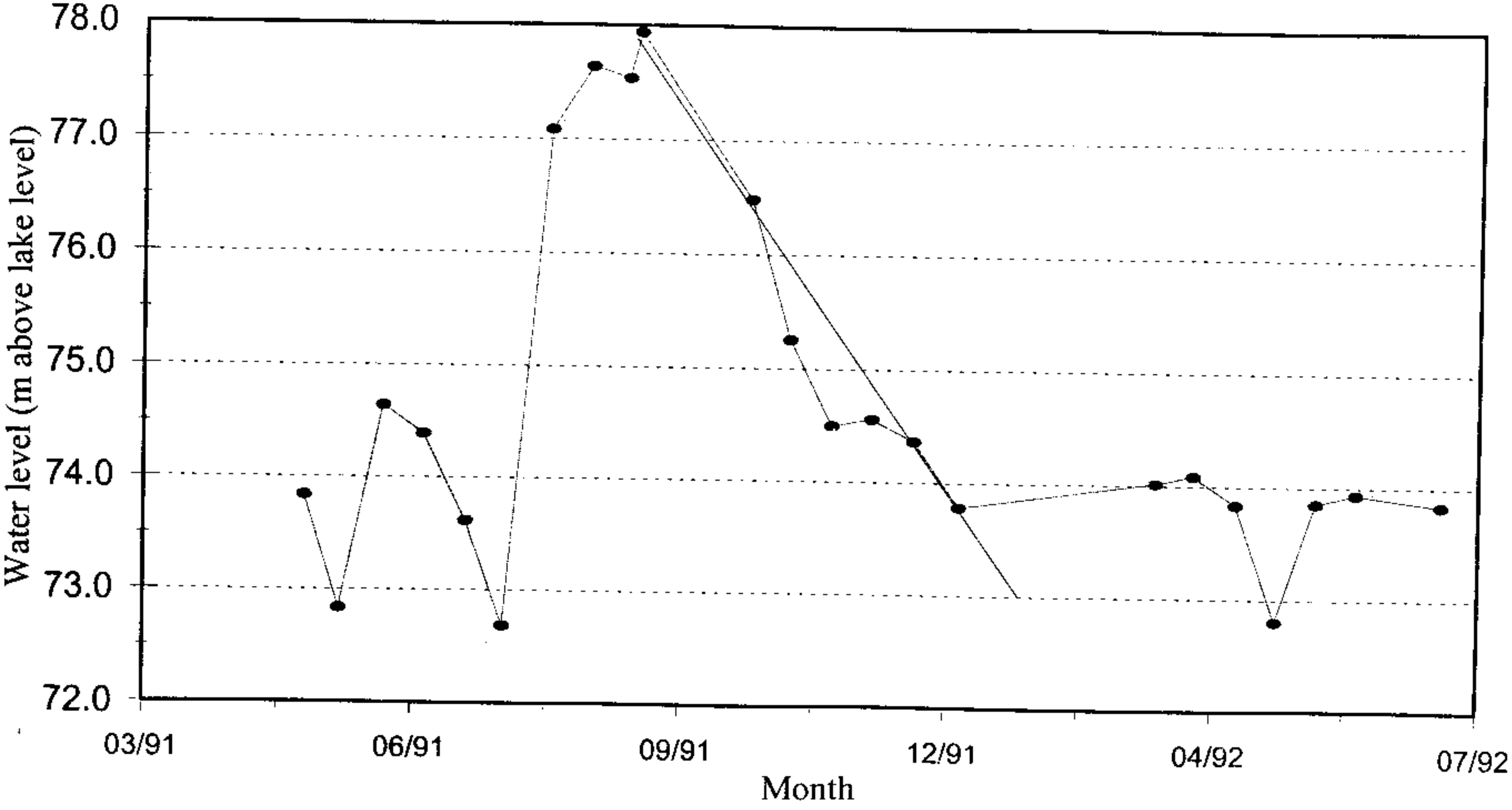
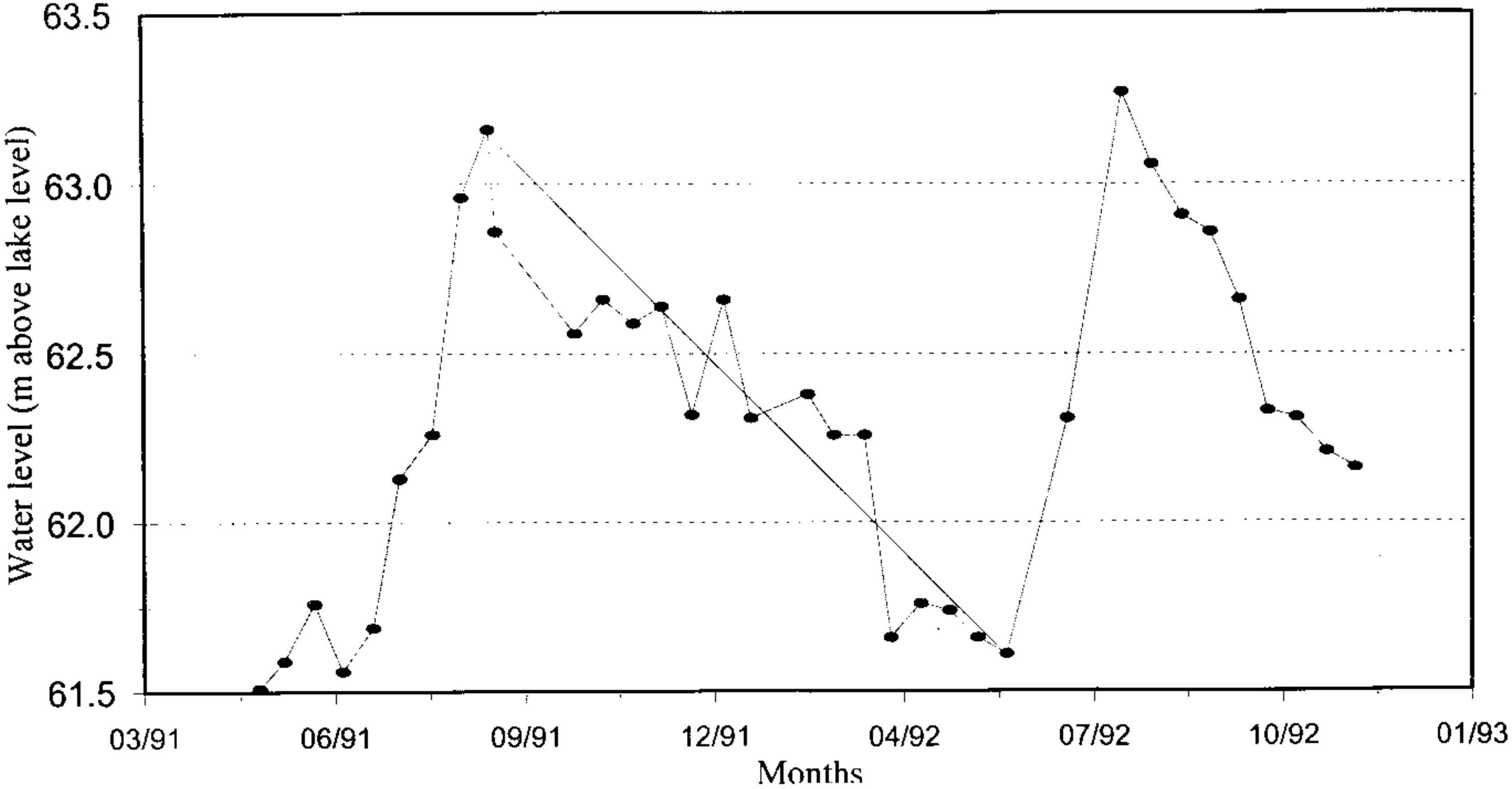


Figure 3.23 Well hydrograph of DH#4 located in Manora Shales above the spring near Spring Cottage, Nainital



From the above discussions it is seen that the Manora shale (as deduced from the crude preliminary tests on the existing observation wells for the aquifer parameters S and T) is an aquifer with very low or poor productivity. Any groundwater movement should occur only through lineaments such as fractures and faults, unfortunately for which no data is available. Further, all the four observation wells installed in the year 1976 are not available any more for further observations of groundwater level (even the well DH#2, in which the slug test was conducted at the start of study, has been rendered unusable by the locals).

3.5.2 Springs:

There are 11 springs that issue from the Manora shales characterising the Sher-ka-Danda - Snowview - Alma Hill slopes in the upstream side of the Lake Nainital. All the 11 springs are associated with major fractures. Based on the discharge data of these springs collected by UPPWD during 1976-79, the springs could be divided into two groups as given in Table 3.28.

Table 3.28 Classification of upstream springs based on average discharge and coefficient of variation in the discharge.

Group No.	Spring	Elevation (m.a.m.s.l.)	Discharge (L/min.)		Coefficient of Variation
			Maximum	Mean	
Group I	Spring Cottage	2015	35	25	36
	Lakeview	2040	30	21	56
	Drain #14	2075	28	20	62
Group II	Oak Lodge	2050	6	4	73
	Drain #3	-	7	5	77
	Alma Spring	2130	30	21	78
	Kumaun Lodge	2060	6	4	82
	Doctor House	1960	6	4	83
	Chunadhara	2035	5	4	85

The classification of the springs into two groups have been made on the basis of Coefficient of variation (CV) of the monthly discharge to the mean monthly discharge. C.V. is defined as the ratio of standard deviation and expressed as percentage (Schuster and White, 1971). Based on the monthly discharge rates, qualitatively the springs may be classified into two groups viz. those that show smaller variations (CV < 70%) as Group I and those show comparatively larger variations (CV > 70%) as Group II. It is also notable that all the springs that fall in Group I have a fairly larger mean monthly discharges (> 20 L/min.) than those of Group II (< 5L/min.), with an exception of Alma Spring. The difference in the discharge rates could be due to the differences in the volumes of the groundwater reservoirs that contribute to the springs.

The largest spring within the lake basin in terms of discharge is the Pardhadhara, that issues near the Mallital Water Works. The Pardhadhara (Dhara=Spring) is also located in the Manora shales but in the vicinity of the junction of Lake Fault and Sleepy Hallow and Snowdon Faults, very close to the Barapathar argillaceous limestone member. Valdiya and Bartarya(1988), however, consider it to be a karst spring related to the underground channels of dolomite cut by the Lake Fault. The C.V., mean & maximum discharge of Pardhadhara are 53, 1540 L/min. and 2613 L/min., respectively, based on the data available for the period 1982-86 with the Kumaun Jal Sansthan. The sheer size of the discharge of the Pardhadhara, in comparison to the other springs discussed previously, distinguishes it from other springs in the basin. Probably part of the groundwater reservoir that is replenished by water that infiltrates in the Sukhatal catchment and Lake Sukhatal is discharged through Pardhadhara.

In general the springs occurring within the Lake Nainital basin may be classified following the procedure adopted by Shuster and White (1971) into two end-member types viz. a) diffuse flow springs along joints, fractures, partings, bedding planes and other small interconnected openings measuring in centimetres or less and b) flow through integrated conduit systems with water flowing, often turbulently, through solution passages measuring from centimetres to meters. The characteristic of these two types have been outlined by Shuster and White (1971). Accordingly the first type of springs tend to have a laminar flow in a porous anisotropic medium, with usually a well defined water table and uniformly distributed hydraulic conductivity. Natural discharge from such a system is usually through a large number of springs and seeps or a few large springs due to stratigraphic or structural controls. The second type, on the other hand, is a conduit system the flow through which approximates the flow through irregular pipes, with localised groundwater flow paths (due to solutional modification). Flow through the conduit system is often turbulent and rocks of very low hydraulic conductivity are found near the rock formation hosting the conduits. The springs occurring in the Manora shales (to the east and northeast of the lake) may be classified as diffuse springs, while Pardhadhara may be classified as conduit spring, as they fulfill some of the characteristics outlined above.

In the downstream side of the Lake Nainital, outside the lake basin, there are about 15 springs. They are S2, S3, and S4 that issue in the Kailakhan area to east of the lake basin, Sipahidhara, Gupha Mahadev Temple Spring, Rais Hotel Spring and Krishnapur Spring that issue in the Balia ravine area to the south of lake basin, Takula, S6, S7, S8, S9, S10, S11 and S12 that issue in the Ruinsi village - Nihal nadi area to the west of the lake basin. Of these 15 springs, only those springs that are located in the Balia ravine are of significance, due to comparatively larger discharge rates. The discharge data for some of the downstream springs are available for the period 1948-52 with the UPPWD. During the current study the discharge rates of Sipahidhara and Gupha Mahadev Temple Springs were monitored. The brief results are presented in Table 3.29.

Table 3.29 Classification of downstream springs based on average discharge and coefficient of variation in the discharge.

Spring	Elevation (m.a.m.s.l.)	Discharge (M^3/day)		CV	Remarks
		Mean	Max		
Dhobighat	1850	160	273	25	Data pertaining to the period 1948-52.
Patwadanger	Not known	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
GM Temple	1785				

It is noticeable that the discharge of Sipahidhara spring has reduced considerably between 1948-52 an 1995. Further, some of the springs (such as Fairy Hall Spring) for which historical discharge data are available have since gone dry.

3.5.3 Groundwater withdrawal

The water supply from springs and lake through pipe lines was first introduced in the year 1898 in Nainital town. During that period springs were tapped directly for domestic use by the general public, which is still continued in some places, while the steam engine plant was used to pump the water to high zones. In 1923, electric pumping stations were installed to supply water from lake to Nainital town. Presently, some deep tube wells as well as an open well, located at the banks of the lake, are being used to pump the lake water including groundwater. There are three tube wells in addition to the eight pumping sets installed in the open well located near in the Mallital bank of the lake. Out of the tube wells, one deep tube well, constructed recently is operated by UP Jal Nigam and the rest are operated by Jal Sansthan, Nainital. The Jal Sansthan and Jal Nigam authorities maintain the records of total pumping hours of tube wells. Therefore, the data of total pumping hours and discharge capacity of pumps & tube wells have been collected from the Jal Sansthan and Jal Nigam offices located at Nainital. One more well is operated by Army authorities at the Mall road side bank of the lake near the Library, to meet the water requirement of military establishment at Lariakantha. The details of the of the discharge capacities of the pumps installed at lake banks are given in Table 3.30.

The total discharge of lake water through pumping has been calculated for the period from Jan. 1994 to Dec. 1995, on monthly basis. The withdrawal of lake water to meet the demand of the Nainital town has been determined by taking 80% efficiency of the electric pumping sets as per the discussion held with the engineers concerned of Jal Sansthan.

Table 3.30 Discharge capacities of wells located in the bank of Lake Nainital

Pump No.	Location	Year of installation	Pump Capacity m^3 / hour
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	Military Pump House	1980	360.00

The monthly data on water pumped from the lake area during the years 94 and 95 is shown in Table 3.31. The monthly discharge of these pumps and tube wells varied between 2,23,580 m^3/month to 4,99,100 m^3/month during 1994-95. The quantum of water pumped varies significantly in different months depending upon the demand. During April to June, pumping is higher and during September to November it is lower (Table 3.31).

Since the tube wells as well as the open well are located on the banks of the lake (5 - 10 m from the lake waters) that are characterised by land slide debris and lake and nala deposits that have higher permeabilities, it is natural that most of the water being pumped is replenished by seepage from the lake.

Table 3.31 Quantity of water pumped (m^3) out from the banks of Lake Nainital in different months during the study period.

Months	1994	1995
January	2,44,470	3,72,970
February	2,43,550	3,56,250
March	2,95,500	4,09,180
April	3,34,900	4,79,270
May	4,59,290	4,99,100
June	4,69,130	4,68,240
July	4,02,760	4,40,690
August	2,68,120	3,21,450
September	2,23,580	2,42,750
October	3,18,820	2,34,330
November	3,34,170	2,95,700
December	3,40,930	3,31,990
Total	39,35,220	44,51,920

Therefore, the water being pumped may be considered as a mixture of groundwater and lake water. The chemical and isotopic characteristics may be used to estimate the relative contribution of lake seepage to the pumping, considering the pumped water as admixture.

3.5.4 Chemistry of groundwater

The groundwater occurring in the Nainital basin may be grouped into three major categories based on the influence of the chemistry of the Lake Nainital. They are a) upstream springs that are uninfluenced by the Lake Nainital, b) the wells located in the northern bank of the Lake that may represent a mixture of groundwater and that of the lake water seepage and c) the downstream springs that may or may not be connected to the lake. In the absence of information on the groundwater table contours, it may be assumed that the groundwater flow direction in the area follows the hydraulic gradient that generally parallels the local topography. This assumption may be verified by the information on the groundwater chemistry in conjunction with that of the Lake waters. Therefore, the groundwater sampling program was designed accordingly. The samples from the upstream springs viz. Lakeview Spring, Pardhadhara and the group of smaller springs viz. Alma Spring, Doctor House Spring and Chunadhara were collected for studying the chemical quality. The samples were collected from the springs as close to the openings as possible, except for Pardhadhara for which the samples were collected from construction outlets. A masonry wall had been constructed by the local authorities, so as to facilitate water supply to the local population, through specific outflow pipes. Similarly, samples were collected from Gupha Mahadev Temple and Sipahidhara springs. Samples were also collected from the springs S2, S3 and S4 that issue in the Kailakhan area, to the east of the Lake basin. The samples were analysed using the procedures given in section 3.4.5.1. The brief results are as given in Tables 3.32a to 3.32g.

Table 3.32a Chemical characteristics of Pardhadhara Spring

Pardhadhara	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	μS/cm	4	520.0	850.0	740	120	16
T. Hardness	mg/L	9	304.0	478.0	409	54	13
Ca^{2+}	mg/L	9	28.9	85.8	58	18	31
Mg^{2+}	mg/L	9	52.0	72.4	64.5	6.2	10
Na^{+}	mg/L	7	6.7	20.0	13.3	4.4	33
K^{+}	mg/L	7	2.9	4.5	3.6	0.5	15
$HCO3^{-}$	mg/L	8	150.0	296.0	255	45	17
$SO4^{2-}$	mg/L	8	88.0	177.0	143	29	20
Cl^{-}	mg/L	8	6.0	26.0	16	6	40

Table 3.32b Chemical characteristics of Lakeview Spring

Lakeview	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	μS/cm	5	630.0	1050.0	830	130	16
T. Hardness	mg/L	9	356.0	555.0	489	62	13
Ca^{2+}	mg/L	9	29.0	97.0	58	20	35
Mg^{2+}	mg/L	9	55.0	103.0	81	16	19
Na^+	mg/L	7	5.0	14.0	10	4	37
K^-	mg/L	7	2.0	5.0	3	1	32
$HCO3^-$	mg/L	9	250.0	392.0	331	46	14
$SO4^{2-}$	mg/L	8	100.0	184.0	149	26	18
Cl^-	mg/L	9	2.0	23.0	16	7	43

Table 3.32c Chemical characteristics of Doctor House / Chunadhara Springs

Doctor House/ Chunadhara	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	μS/cm	3	630.0	940.0	740	140	19
T. Hardness	mg/L	8	280.0	498.0	386	83.8	22
Ca^{2+}	mg/L	7	32.1	40.1	35.4	3.1	9
Mg^{2+}	mg/L	7	47.6	100.6	68.5	18.5	27
Na^+	mg/L	7	8.0	22.3	12.4	4.4	35
K^-	mg/L	7	2.2	7.8	5	2.3	45
$HCO3^-$	mg/L	8	200.0	360.0	266	60.1	23
$SO4^{2-}$	mg/L	8	69.7	171.0	124	40.4	33
Cl^-	mg/L	8	18.0	28.0	22.6	2.9	13

Table 3.32d Chemical characteristics of Sariyatal Spring

Sariyatal	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	µS/cm	4	590.0	800.0	670	87	13
T. Hardness	mg/L	8	290.0	419.0	357	39	11
Ca^{2+}	mg/L	8	31.0	73.0	49	13	26
Mg^{2+}	mg/L	8	40.0	67.0	57	8	14
Na^{+}	mg/L	6	0.9	3.8	2	1	48
K^{+}	mg/L	6	1.5	2.6	1.9	0.4	22
$HCO3^{-}$	mg/L	7	217.0	286.0	250	22	9
$SO4^{2-}$	mg/L	8	82.0	150.0	103	22	21
Cl^{-}	mg/L	8	5.0	30.0	15	8	52

Table 3.32e Chemical characteristics of Springs located in Kailakhan area

Kailakhan springs S2 / S3 / S4	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	µS/cm	3	480.0	890.0	630	180	29
T. Hardness	mg/L	8	172.0	398.0	268	75	26
Ca^{2+}	mg/L	8	31.0	63.0	44	12	26
Mg^{2+}	mg/L	8	23.0	62.0	43	13	31
Na^{+}	mg/L	8	10.0	32.0	17	9	54
K^{+}	mg/L	8	2.0	3.0	2	0.5	22
$HCO3^{-}$	mg/L	7	74.0	254.0	145	69	48
$SO4^{2-}$	mg/L	8	106.0	208.0	147	34	23
Cl^{-}	mg/L	8	11.0	25.0	16	5	29

Table 3.32f Chemical characteristics of Sipahidhara Spring

Sipahidhara	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	μS/cm	6	430.0	790.0	630	110	17
T. Hardness	mg/L	11	302.0	370.0	330	20	6
Ca^{2+}	mg/L	11	36.0	68.0	46	8	18
Mg^{2+}	mg/L	11	43.0	61.0	53	4	8
Na^{+}	mg/L	9	5.0	20.0	10	5	44
K^{+}	mg/L	9	3.0	6.0	4	1	26
$HCO3^{-}$	mg/L	11	218.0	288.0	239	18	7
$SO4^{2-}$	mg/L	10	81.0	150.0	106	23	22
Cl^{-}	mg/L	10	8.0	26.0	16	5	33

Table 3.32g Chemical characteristics of Gupha Mahadev Temple Spring

Gupha Mahadev Temple	units	Number of samples	Range		Mean	σ	C.V.
			Min	Max			
E.C.	μS/cm	5	470.0	800.0	620	110	18
T. Hardness	mg/L	10	246.0	356.0	322	31	10
Ca^{2+}	mg/L	10	32.0	51.0	43	7	16
Mg^{2+}	mg/L	10	39.0	58.0	53	5	10
Na^{+}	mg/L	8	4.0	14.0	9	3	37
K^{+}	mg/L	8	1.2	3.3	2.6	0.6	23
$HCO3^{-}$	mg/L	10	159.0	280.0	228	28	12
$SO4^{2-}$	mg/L	9	75.0	133.0	98	21	22
Cl^{-}	mg/L	9	9	30.0	18.0	7	42

It is well known that shales chiefly contain residual minerals cemented usually by calcium carbonates, silica, clay or iron oxides. The principal soluble material in shales rocks are calcium carbonates, and magnesium carbonates. Therefore, one may expect Ca and Mg ions to dominate. The chemistry of Lakeview spring and Chunadhara & Doctor House springs reflect these general characteristics. However, the extent of seasonal variation of the concentration of different ions in these springs (given by the C.V.) probably reflects the volume of the reservoirs they represent. Due to probable shorter residence time the Doctor House and Chunadhara Springs exhibit relatively larger C.V. for most of the major ions that were analysed, in comparison to those of the Lakeview Spring.

Table 3.33 Comparison of average Total Hardness and CMR along with Coefficients of variation for different springs

	Total Hardness (mg/L)		Ca^{2+} (epm) / Mg^{2+} (epm)	
	Mean	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
Lake Nainital South basin Epilimnion	321	16	0.49	18
Sipahidhara Spring	328	6	0.49	13
Gupha Mahadev Temple Spr.	320	10	0.48	13

In case of carbonate rocks, the proportion of calcium to magnesium may reflect the range of Ca:Mg ratio of the rocks. The molar Ca/Mg ratios (CMR) in groundwater have both hydrologic and geochemical significance. In calcitic rocks the CMRs are generally greater than one, and in case of pure dolomites they approach a value of one as seen from standard literatures (Appello, 1988; Appello and Postma, 1997; Fetter, 1997). However, the CMRs for all the spring waters in the study area, are less than one in all the seasons (Table 3.33). This condition that

CMR < 1.0, in groundwater may be brought about by: a) introduction of substances with CMR < 1.0 into the groundwater in the recharge areas, b) phosphorous in groundwater recharge and c) incongruent solution of dolomite with calcite precipitation (Langmuir, 1971).

The possibility of the first case being the cause for low CMR does not exist in the study area, as no such activity - for treatment of top soil - is carried out. The second case i.e. phosphorous in recharge is crucial as calcium phosphate compounds are less soluble than magnesium compounds (McCarty et al., 1970). The possibility of phosphorous in groundwater recharge includes use of fertilisers, animal wastes, phosphate detergents, leakage of septic tanks and sewage effluent discharge facilities. Among these activities the use of fertilisers and animal wastes may be ruled out as there are no agricultural or farm activity in the lake basin. However, the chances of pollution do exist as the population in the lake basin is substantially high (40,000 during 1991 census). This factor could be tested by analysing the hydrochemical data of different seasons pertaining to the springs for significant correlation between CMR and Na^+, as phosphate in the human waste or sewerage is associated with Na^+. Since there is no significant correlation between CMR and Na^+ this possibility may also be excluded from further consideration.

The chemical analyses data of samples collected from the springs during summer of 1995, when compared to those collected during winter of 1995 exhibit an increase of 50% in Ca^{2+}, 51% in CMR and 2.7% in Mg. Further, the significant positive correlation (0.82 - 0.98) between CMR and Ca^{2+} (much higher than with Mg^{2+}) in all the upstream springs (considering the data pertaining to the samples collected during different months between March, 1995 and April, 1996) and the lower coefficient of variation for Mg^{2+} (7 - 16) compared to Ca^{2+} (~30) indicate that the variation in CMR is mainly due to the variation in Ca^{2+}. The above properties of the Nainital basin groundwater indicate that between winter (influenced by post-monsoon effects) and summer, precipitation of calcite must be occurring concurrently with dissolution of dolomite. The incongruent dissolution of dolomite may be explained with the "Carbonate solution and precipitation model" proposed by Langmuir (1971).

Langmuir (1971), in a study on the geochemistry of groundwater of Central Pennsylvania (U.S.), the characteristics of which is not very different from that in the Nainital basin, postulated that the low CMR in groundwater of carbonate terrain could be brought about by the incongruent dissolution of dolomites. The model presented by Langmuir (1971) a groundwater in near saturation with the carbonates may have the chemical characteristics that is related to the changes in CO_2 pressure or the changes in groundwater chemistry itself that influences the P_{CO_2}. As the CO_2 in groundwater increases, more carbonate rocks dissolve such that there is an increase in the HCO_3 and decrease in the pH. Similarly, an increase in pH, Ca^{2+} or HCO_3 can cause calcite precipitation with a consequent increase in P_{CO2}. This characteristic is exhibited by the groundwater of Nainital basin as evidenced by the negative correlation between HCO_3 and pH (-0.86) for the summer samples and also by the lower P_{CO2} during winter (-2.91) than summer (-2.45).

Langmuir (1971) also recalculated the activity product of dolomites (K_d) in Central Pennsylvania based on the mean observed CMR in the groundwater (~0.6) and showed that the groundwater would be more aggressive (undersaturated) with the Saturation Index (SId) corresponding to the dolomites of the study area than for ideal dolomites. In the case of Nainital groundwater the CMR is still lower (~0.5). Therefore, the K_d for the natural dolomites in the basin will be slightly higher than for ideal dolomites and hence the SId values will be lower than the values presented in Table 3.34. The SIc, SId, and SIg values presented in the Table 3.34 have been calculated using the WATEQ program, that assumes Kd value of 10^{-17} at 25°C against a higher value of $10^{-16.72}$. This value has been calculated using the CMR values presented by Langmuir (1971) for different temperatures considering the CMR for both ideal dolomite and also the natural dolomites of Central Pennsylvania. The calculation of Kd from CMR is based on the following equation (Holland et al., 1964):

$$\frac{(K_c)^2}{K_d} = \frac{[Ca^{2+}]}{[Mg^{2+}]}$$

The K_c values for different temperatures are also presented by Langmuir (1971). Since, the groundwater at Nainital at 15°C have a CMR of ~0.5, the K_d value at 15°C is $10^{-16.44}$ as calculated from the above equation. Using the temperature - CMR relationship presented by Langmuir (1971) the CMR at 25°C is ~0.83 and the corresponding K_d value is $10^{-16.72}$. If this higher value for K_d is used in place of K_d for ideal dolomite (16.85 at 15°C) then the SId values given in Table 3.34 are lesser by 0.1. Therefore the carbonate rocks of Nainital have higher solubility than ideal dolomites, because of mixed layering (Figure 3.1).

Table 3.34 Saturation indices of Calcite and Dolomite of the spring waters sampled during different seasons in 1995.

Season	Spring ID	SIc	SId
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

Further, as noted by Hem (1985) magnesium occurs in significant amounts in most limestones, and the dissolution of the rock results in dissolution of magnesium, but the process is not ordinarily reversible, i.e., the precipitate that is formed from the solution is usually calcite, and therefore the concentration of magnesium would tend to increase along the flow path of groundwater until a rather low CMR is reached. These processes discussed above adequately explain the low CMR found in the groundwater of the Nainital basin.

The Na^+, K^- as well as Cl^- are found in very low concentrations compared to alkaline earth and sulphate ions. The maximum concentration of these ions are 22.3, 7.8, and 30 mg/L for Na^-, K^+, and Cl^-, respectively. However, the relative variation of Cl- is very helpful in estimating the evapotranspiration index - ET_{index} (also called as recharge index). ET_{index} is defined as (Mazor, 1997):

$$ET_{index} = \frac{R_c}{GW_c} \cdot 100 \qquad (3.31)$$

where, Rc is the concentration of Cl^- in rainwater and GWc is that in the groundwater. Expressed in percentage the ET_{index} can be used to estimate the approximate loss due to evapotransipration during the process of groundwater recharge. Assuming an average Cl^- concentration of 1.0 mg/L in the Nainital rainfall (Section 3.2.5) and using the mean observed Cl^- in the groundwater of the Nainital basin (16 mg/L), the average loss due to evapotranspiration is ~93.7%. In other words, the quantum of water infiltrating into the soil surface reaching the zone of saturation is ~6.3% only. Sulphates in the groundwater of Nainital basin may arise from the oxidation of pyrites. The possibility of oxidation of pyrites in the Nainital basin has been noted by several geologists who investigated the basin (Middlemiss, 1890; Coulson, 1939). The oxidation and hydrolysis of pyrite produce sulphuric acid and soluble sulphates as follows (Karanth, 1987):

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

$$FeS_2 + H_2O + 7O \rightarrow 2FeSO_4 + H_2SO_4$$

$$2FeSO_4 + H_2SO_4 + O \rightarrow Fe_2(SO_4)_3 + H_2O$$

Sulphuric acid thus formed reacts with calcite, in the weathered zones and calcareous rocks, to produce soluble calcium sulphate:

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

The nature of slates and dolomites being pyritiferous in the Nainital basin has been reported earlier investigators (Coulson,1939; Ashraf ,1978; and Valdiya,1988). Therefore, the origin of sulphates in the groundwater are probably due to oxidation of pyrites.

3.5.5 Stable isotope ratios

The samples for stable isotope analyses were collected from the springs as well as the open well located on the lake bank, were collected using the procedure given in Section 3.4.5.1. The samples were analysed using the procedure given in Section 3.4.5.1. at the IRMS Laboratory of H & T Section, Isotope Division, BARC, Mumbai. One set of samples collected during September, 1996 were analysed at EIL, University of Waterloo, Ontario, Canada. The results are given in Table 3.35.

Table 3.35 Observed $\delta^{18}O$ values of the 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	Sariatal	-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 seen from the Table 3.35, that both $\delta^{18}O$ and δD vary during different seasons. The variation in $\delta^{18}O$ is larger for the Pardhadhara spring (-7.3‰ during post-monsoon and -8.8‰ during monsoon). The variation is probably due to the presence of Lake Sukhatal, the temporary lakelet in the recharge area of the Pardhadhara spring. The Sukhatal suffers evaporation during the brief period of its existence (June - September) and its water that has undergone evaporative enrichment mixes with the sub-surface reservoir. The $\delta^{18}O$ during the winter of 1994 is slightly depleted than the post-monsoon values of 1995. This probably could be due to the infiltration of the melt of snowfall that occurred during the winter of 1994. However, this could not be verified as the isotopic composition of the snowfall of 1994 is not available. The springs located in the Snowview - Alma Hill - Naina peak slope, viz. Chunadhara, Alma Cottage Spring and Doctor House Spring have $\delta^{18}O$ values of -9.1 to -9.4‰ during monsoon. Compared to these, the $\delta^{18}O$ of the Lakeview spring varies between -8.0‰ and -9.0‰ indicating that the source of the spring is mixed with recharge of enriched water. The post-monsoon $\delta^{18}O$ values of Alma Cottage Spring and Lakeview Spring are comparable (-8.3‰) that is heavier than the value observed during monsoon in case of Alma Cottage Spring. This is indicative of delayed groundwater recharge that has undergone partial evaporative enrichment. The frequency distribution analyses show that for monsoon season the peak is about -9.0‰ and for non-monsoon season the peak is skewed towards -8.2‰. This shows that the post-monsoon and winter recharge to groundwater has a definite overall enrichment effect on the groundwater $\delta^{18}O$ values.

Compared to the rainfall weighted $\delta^{18}O$ for the monsoon season (-11.3‰) the mean monsoon groundwater $\delta^{18}O$ (-9.0‰) shows an enrichment of 2.3‰. This is much higher than the average enrichment of 1.0‰ values reported by Gat and Tzur (1967) for a sandstone formation with no permanent surface water bodies in Israel. Gat and Tzur (1967) also pointed out that the process of interception may lead to an overall enrichment of recharge waters under certain conditions, upto 2.0‰. Further, if the multiple-linear model (Equation 3.4) is used with the mean monthly data pertaining to the months that had a rainfall in excess of 10mm, we get a $\delta^{18}O$ value of -9.0‰. The value is comparable to the mean observed $\delta^{18}O$ value of the local groundwater. This, however may be interesting, has to be used with caution as the said model itself has been developed on the basis of single year rainfall $\delta^{18}O$ data.

From the variation of observed infiltration rates in the catchment (Section 3.3.3) it is possible that the rainfall falling in the zone of higher rate of infiltration percolates and reaches the zone of saturation quickly than the rainfall falling in the zone of lower rate of infiltration. Although the distribution of zones of different rate of infiltration is not well defined in the catchment, the infiltration is in general higher in the linear depressions that generally coincide with the prevailing surface drains. The surfacial sheet flow over the impermeable or less permeable soil cover and the interflow that reaches the zone of saturation may be heavier in $\delta^{18}O$ value than the water that reaches the zone of saturation by direct infiltration and percolation.

However, these different pathways lead to an overall enrichment of the groundwater $\delta^{18}O$ values. Another reason for the isotopic enrichment in groundwater could be due to flushing of soil water that has undergone evaporative enrichment, by the infiltration of fresh rainfall. This effect is more pronounced for the bare top soil than the top soil covered with grass (Zimmermann et al., 1967). Since ~50% of the catchment area of Lake Nainital is characterised by non-forestry landuse (Section 3.3.1) with less or no tree cover it is possible that the soil water suffers evaporative enrichment. The stable isotope characteristics of springs located outside the catchment of Lake Nainital are of significant importance, as they may give insight to the probable hydraulic interconnection between the lake and the sources of the springs. Keeping this in view, almost all the springs that issue outside the lake basin and having the issue points at elevations below that of the lake surface were sampled for $\delta^{18}O$ analyses during December, 1994. The results are given in Table 3.36.

Table 3.36 $\delta^{18}O$ values of downstream springs. Samples collected during December, 1994.

Spring ID	Altitude (m amsl)	$\delta^{18}O$ (‰)	Spring ID	Altitude (m amsl)	$\delta^{18}O$ (‰)
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

Further, the Gupha Mahadev Temple Spring was sampled for stable isotope analyses during different months to establish the nature of variation and zone of the Lake Nainital that may contribute to the discharge of the spring. The results are given in Table 3.37.

Table 3.37 $\delta^{18}O$ values of Gupha Mahadev Temple spring. Samples collected during different seasons.

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

The large variations in the $\delta^{18}O$ values of Gupha Mahadev Temple Spring during different seasons indicate that the source of the spring is highly influenced by a surface water body that undergoes evaporative enrichment.

4.0 LAKE NAINITAL - GROUNDWATER INTERACTION

4.1 Conceptual Model for the Lake Water Balance

The lake water balance approach physically accounts for the components of outflow from the system, inflow to the system and changes of storage within the system. The general equation of water balance of a lake could be written as

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

where, ΔV is the change in storage during a selected time. The different flow components for the conceptual model of the Lake Nainital water balance are shown in Figure 4.1. The water balance method can be used to estimate the sub-surface components viz., the groundwater inflow to and from the lake, if all other flow components are accurately determined. Incorporating the different inflow and outflow components in the above equation and rearranging we get,

$$SSi - SSo = (Eo + So + Wo + \Delta V) - (Pi + Si + Di) \qquad (4.1)$$

Where
SSi = sub-surface inflow to the lake $[L^3/T]$
SSo = sub-surface outflow from the lake $[L^3/T]$
Eo = free water evaporation from the lake surface $[L^3/T]$
Wo = Withdrawal from the lake $[L^3/T]$
ΔV = change in lake storage $[L^3/T]$
Pi = direct precipitation over the lake surface $[L^3/T]$
Si = surface water inflow to the lake $[L^3/T]$
Di = inflow to the lake through the drains $[L^3/T]$

Since there are two unknown components (SSo and SSi) in a single linear equation, the SSo must first be known by having a reliable estimate of the component. This could be achieved by characterising the springs in the downstream area of the lake by non-destructive tracer techniques, that may provide an insight as to identify those springs that are hydraulically connected to the lake.

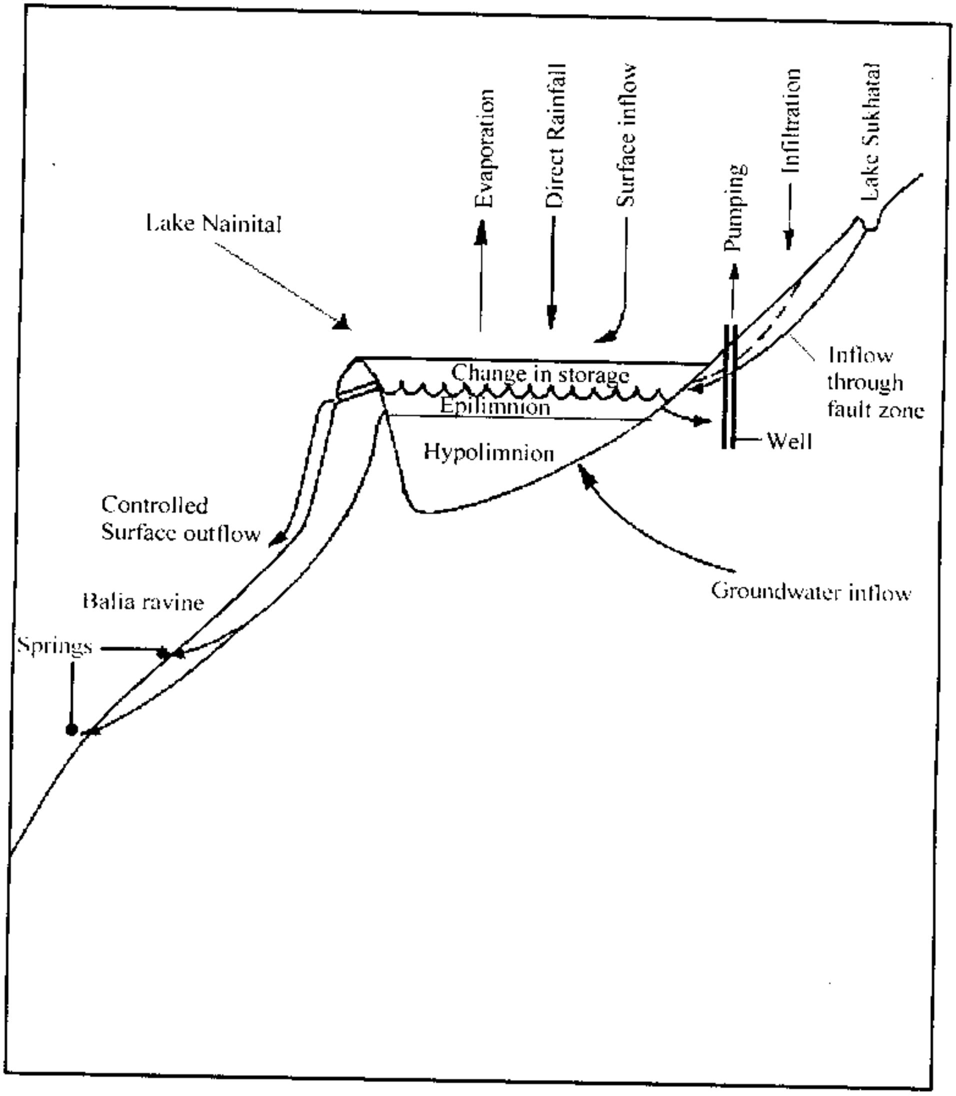


Figure 4.1 Conceptual model showing various water balance components of Nainital lake.

The water balance approach can be used to determine estimates of unmeasured components of the lake system under certain conditions only. Firstly, in unsteady flow conditions the changes in storage of the system occur over a finite interval of time, hence the time interval selected for incorporating into the equation must be large enough to allow for the accounting of the changes in storage of the system. Secondly, in the water balance approach the unknown component is computed as the residual, hence all the known components must be accurately determined / estimated for the selected time interval. Finally, the magnitude of the computed component should be large relative to the sum of the errors of known components; otherwise, the computed component may be masked by the error of the computation.

As Lake Nainital is located at the mouth of the lake basin – unlike most of the natural lakes that are located more centrally – the lake reflects the sum of all the processes that take place within the lake basin. The change in storage of the lake responds quickly to the direct precipitation and surface inflow. The latter component is effected by steep slopes in the Lake Basin and hence higher gradients. Because, the elevation difference between the lake surface and the highest point in the periphery of the lake basin is about 600m and the distance between the highest point and the basin outflow is less than four kilometres, the time of concentration is very less. The available daily lake stage data that are observed at 24-hour interval adequately reflect the changes in lake storage brought about by direct rainfall and surface run-off. The other major component that influences the change in storage is the surface outflow. The outflow from the lake is a controlled one and as described in Section 3.4.4 occurs only during monsoon season. Evaporation from the lake surface during 24-hour time interval is of the order of 1 – 2 mm, and is not readily discernible from daily lake stage data. Apart from the above components that can be measured or estimated using standard methods and established empirical / semi-empirical methods, the lake stage is influenced by withdrawal from the lake.

Since the lake basin is characterised by a multitude of fault and fracture zones that coincide with the 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 not be higher. Due to the nature of the weathered zone, not extending to greater depths, the infiltrated water may move along preferential pathways and find its way to the lake.

The groundwater level gradient towards the lake supports this hypothesis. The coefficient of variation historical monthly spring discharge data and the chemistry of the springs indicate that the groundwater residence time may not be too large. The Sukhatal sub-basin of Nainital does not has any surface outflow. The sub-basin is a closed one. All the water that is received as precipitation in the basin is lost in a short time, through infiltration and evaporation. Because of the proximity of the Lake fault to the Sukhatal Lake, it is possible that most of the water lost to underground seepage recharge the Nainital Lake.

The withdrawal from the lake does not take place from lake proper, but from the wells installed at the periphery of the lake. Assumption that the entire quantum of water that is pumped from these wells is seepage from the lake may be misleading. This has to be verified by tracer techniques. However, the wells are located very close to the lake and in a formation characterised by young unconsolidated landslide debris and lake sediments that may have higher permeability. Therefore, a major portion of the pumped water is probably replenished by subsurface outflow from the lake.

The sub-surface outflow towards the downstream side of the Lake may be effected by the presence of fractures and faults in the geological formations. The seepage from the lake may not be occurring through the lakebeds as they are characterised by thick layer of fine sediments and hence any sub-surface outflow must be occurring in the epilimnion and/or mesolimnion zones. This outflow from the lake then recharges the unconfined aquifer that in turn discharges through the numerous springs located in the downstream side of the lake. Therefore, in the absence of information on the groundwater levels in the unconfined aquifers of the downstream side, the seepage from the lake could be quantified through the discharge measurements of those springs that are hydraulically connected to the lake. The investigations for the verification of lake - spring interconnection can be made by using tracer techniques.

4.2 Interconnection of Lake Nainital and Downstream Springs

The chemistry of water sampled from the lake Nainital and the springs, the possible sources of different ions and the controlling factors have been discussed in detail in Sections 3.4.5 and 3.5.4. When sulphate concentrations are plotted against the more conservative chlorine (Figure 4.2), the values of Lake Nainital and the few downstream springs viz. Sariyatal and Balia ravine springs plot in a single cluster. The other downstream springs such as S9, S12, Kailakhan Springs (S2, S3 & S4) and the Durgapur Spring do not fall in the lake cluster. This may be indicative of the fact that the Balia Ravine springs and Sariyatal spring may be connected to the lake.

In order to complement the information obtained through the hydrochemistry, stable isotopic investigations were carried out (Table 3.36). During winter, when the lake is well mixed, the Gupha Mahadev spring shows an $\delta^{18}O$ value (–9.5‰) close to that of the Lake (-9.6‰). However, the springs S2, S3 and S4 in the Kailakhan area show a comparatively heavier $\delta^{18}O$ (-7.0 to -7.5 ‰). Since, the lake has an $\delta^{18}O$ value of about -7 ‰ during July / August, the enriched values in the Kailakhan springs suggest the possibility of leakage from the lake and a travel time of 4-5 months. If the waters that issue from springs S3 and S4 are considered as leakage from the Lake Nainital, then the change in water type from magnesium bicarbonate to magnesium sulphate could be ascribed to the long travel time and contact with the highly shattered pyritiferous shales.

The springs S1, S6, S7, S9, S10, S11 and S12 are not connected to the Lake, as their $\delta^{18}O$ values (-10.2 ‰ to –11.8 ‰) are depleted than that of the lake (-5 ‰ to –9.6 ‰). The inference that Sipahidhara spring is connected to the Lake through lineaments is substantiated by the chemical concentrations. When total cations (TZ^+) values of different springs are normalised to that of Lake Nainital (Figure 4.3), Sipahidhara spring shows little variation with time in comparison to the upstream springs and wells located with in the lake basin. This indicates that the lake is the main source for the Sipahidhara spring and the local groundwater has little influence on the chemistry of the spring.

Fig. 4.2 Sulfate – Chloride cross plot for different water sources

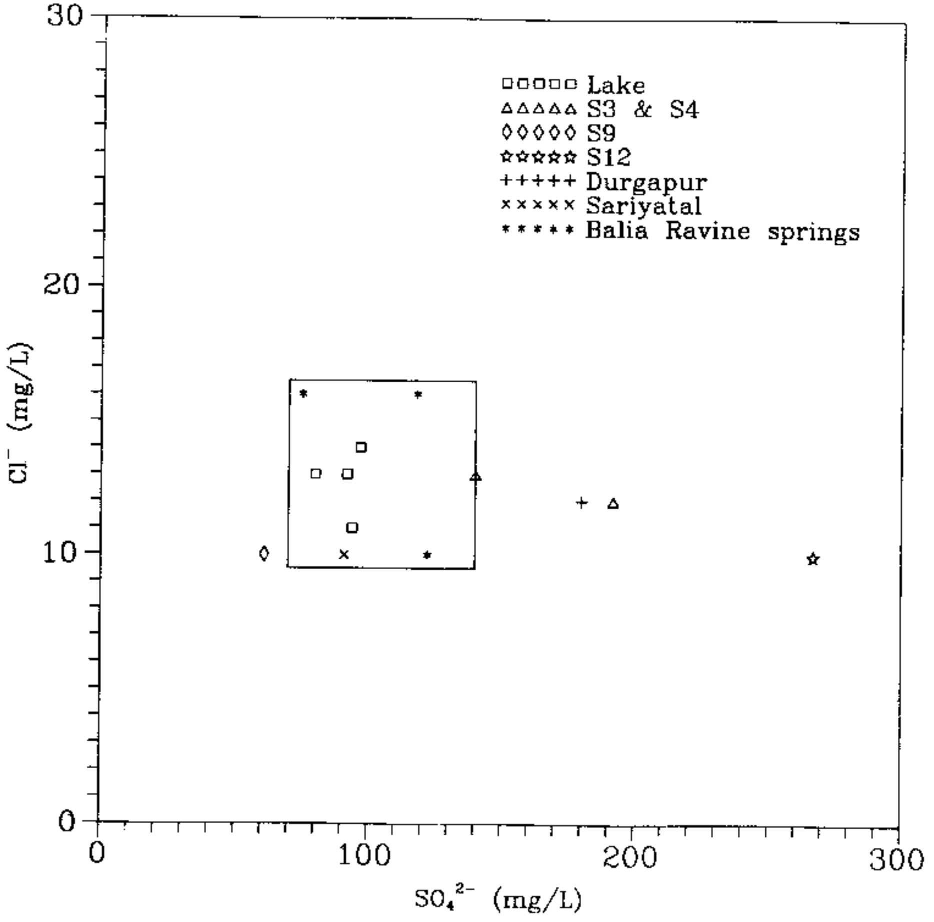
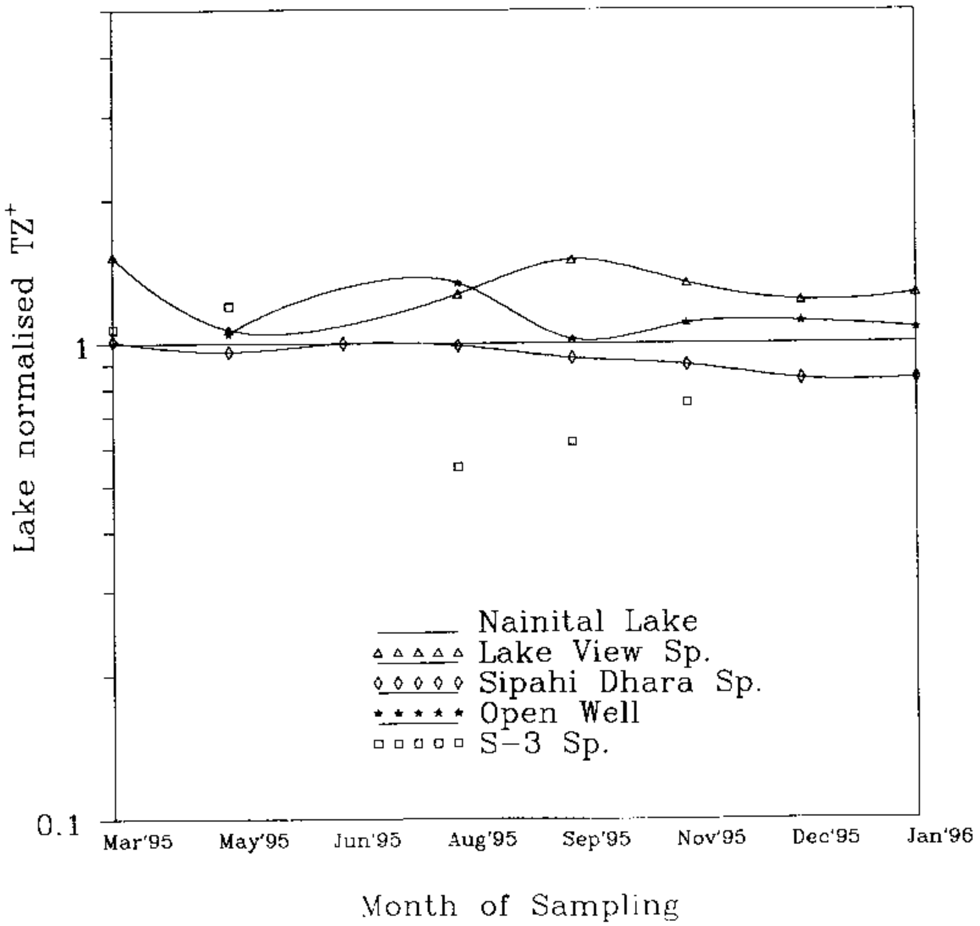


Fig. 4.3 Variation of total cations (TZ^+) of different water sources with time



Comparison of ion concentrations of the series of springs (around Sipahidhara and Gupha Mahadev located in the Balia ravine) to the vertical concentration profile of the lake, indicates that these springs receive water from the Lake's epilimnion / mesolimnion zone. This is also substantiated by the oxygen isotope data pertaining to Lake Nainital and Gupha Mahadev Spring presented in Tables 3.26 and 3.37, respectively. The pattern of variation in the $\delta^{18}O$ values of the Epilimnion of the lake Nainital and that of the Gupha Mahadev Spring are similar as seen from Figure 4.4.

4.3 Two-Component Mixing Model for the Estimation of Proportion of Lake Water in the Pumpage (Wo)

As outlined in Section 4.1, the proportion of the lake water in the water pumped from the wells located in the periphery of Lake Nainital can be estimated by tracer techniques. The two-component mixing model is highly useful in such conditions. The groundwater isotope index has been calculated from the data pertaining to the springs located upstream of the lake (Table 3.35), by taking the average values of different springs. The isotopic index of the lake has been computed from the data presented in Table 3.26, and the volume-weighted values have been used by considering the volume of epilimnion and hypolimnion zones. The oxygen isotope data of the admixture i.e., the well waters along with the results are presented in Table 4.1. 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 4.1 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

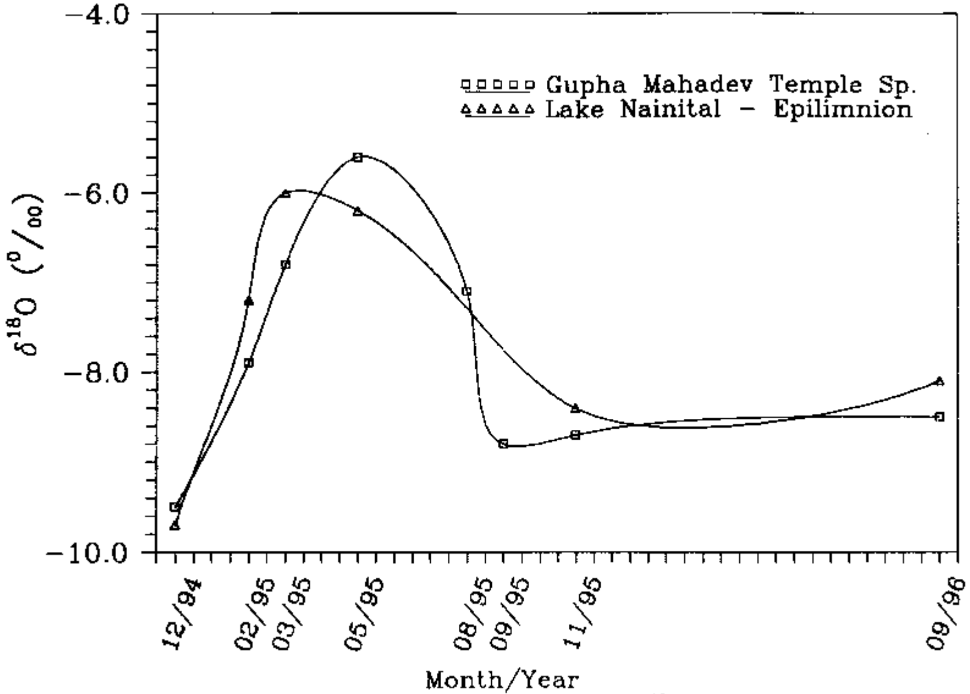


Figure 4.4 Temporal variation in $\delta^{18}O$ of Gupha Mahadev Spring and Lake epilimnion zone

4.4 Estimation of Sub-Surface Outflow Through Springs

Average monthly discharge of the downstream springs located in the Balia ravine were monitored by the PWD for few years (1947-52). The spring discharge data are very useful in arriving at an approximate estimate of sub-surface outflow. Out of the nine springs such as, Tallital spring, Fairy Hall Spring, Old Reserve Police Lines (Rais Hotel), Mota Pani (Sipahidhara), Katarpani, Cantonment drain etc., Rais Hotel and Sipahidhara alone account for 90 - 94% of the total annual spring discharge. Some of the springs for which discharge were measured have since gone dry, and the discharge of major springs such as Sipahidhara has decreased considerably. The aggregate monthly discharge of all springs and that of Sipahidhara have been presented in Table 4.3. Due to logistic constraints, the discharge of Sipahidhara spring alone was monitored during 1995. Based on the mean monthly ratio of discharge of Sipahidhara spring to the total discharge of all springs using the available historic data, the total discharge of all the springs that are connected to the Lake for the year 1995 has been computed.

Table 4.3 Comparison of the discharge of Sipahidhara spring (SD) and that of all springs that were measured during the period 1947 to 1952 and 1995. (Values given are in thousands of cubic meters.)

Year	1947		1948		1949		1950		1951		1952		1995
	SD	Total	SD	Total	SD	Total	SD	Total	SD	Total	SD	Total	SD
Jan	94	169	88	165	135	260	88	171	87	178	87	178	14
Feb.	79	143	79	148	109	199	80	156	79	156	79	156	11
Mar.	91	163	89	169	108	195	90	171	89	172	89	172	1
Apr.	87	156	84	159	83	160	80	157	81	152	81	152	0.60
May	91	163	75	150	75	138	75	146	33	89	33	87	-
Jun.	95	169	63	125	73	142	71	140	31	82	30	82	-
Jul.	98	181	70	154	300	539	-	36	302	424	109	223	-
Aug.	94	178	301	554	301	549	401	709	401	583	201	361	40
Sep.	94	171	315	588	389	671	291	535	290	530	146	316	50
Oct.	95	172	302	570	241	446	235	443	235	461	139	293	28
Nov.	88	162	246	425	166	327	164	328	164	324	124	253	19
Dec.	91	166	172	309	152	256	156	263	155	260	116	233	18

From the data presented in the above table, it is seen that there is a reduction of about 85% in the discharge of the Sipahidhara spring.

4.5 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 (Winter, 1981). 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 4.4.

Table 4.4 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%	Winter, 1981
Surface Inflow	S_i	SCS-CN and LLTA methods	20%	
Inflow through drains	D_i	Discharge measurements at discrete time intervals	15%	Winter, 1981
Evaporation loss	E_o	Modified Penman method	15%	Winter, 1981
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 (4.3)

4.6 Water balance for the Lake Nainital

The estimated / calculated values of each component of the water balance for the period January, 1994 to December 1995 has been presented in Table 4.5. along with the standard error in the estimates. The groundwater inflow to the lake using the water balance approach has been computed using the following equation that has been derived by rearranging equationn (4.1).

$$SSi = (Eo + So + Wo + SSo \pm \Delta V) - (Pi + Si + Di) \qquad (4.2)$$

The error in the estimates of SSi has been computed using the following expression:

$$\sigma_{SSi} = [\sigma_{Eo}^2 + \sigma_{So}^2 + \sigma_{Wo}^2 + \sigma_{SSo}^2 + \sigma_{\Delta V}^2 + \sigma_{Pi}^2 + \sigma_{Si}^2 + \sigma_{Di}^2]^{1/2} \qquad (4.3)$$

It is seen from the Table 4.5, that the absolute error in the computed values are relatively less compared to the computed values. The errors do not mask the results (except for the months of April and June 1994) and therefore the results may be considered as reliable. The errors in the computed values are larger for the monsoon period, which is due mainly to the larger error in the surface outflow component. It is also seen that the error relative to the computed sub-surface inflow is larger.

4.7 Isotope Mass Balance

The isotope mass balance of a lake may be written as

$$\Delta \delta V = (\delta_P Pi + \delta_{Si} Si + \delta_{Di} Di + \delta_g SSi) - (\delta_E Eo + \delta_{So} So + \delta_L SSo) \qquad (4.4)$$

The equation could be rearranged to get the net of the sub-surface terms:

$$\delta_g SSi - \delta_L SSo = (\delta_E Eo + \delta_{So} So \pm \Delta \delta V) - (\delta_P Pi + \delta_{Si} Si + \delta_{Di} Di) \qquad (4.5)$$

where SSi, SSo, Eo, So, Pi, Si, and Di are as given in eqn. (4.1) and δ_g, δ_{Go}, δ_E, δ_{So}, δ_P, δ_{Si}, and δ_{Di} are the corresponding isotopic values expressed in permil with respect to the V-SMOW standard. Solving equations (4.4) and (4.5) simultaneously, we get (simplified):

$$SSo = [\delta_g (O \pm \Delta V - I) - (\delta_O O \pm \Delta \delta V - \delta_I I)] / (\delta_{Go} - \delta_g) \tag{4.6}$$

The above equation permits independent determination of the sub-surface outflow component of the lake, which in turn could be used to estimate the groundwater inflow to the lake using the following relation:

$$SSi = [(Eo + So \pm \Delta V) - (Pi + Di + Si)] + SSo \tag{4.7}$$

Table 4.5a 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 4.5b 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

This method eliminates the need to estimate the outflow through pumping or discharge through springs from the lake. Isotope mass balance has been attempted for the period between February, 1994 and February 1995. Since in the month of February the lake is well mixed and homogenous, it eliminates the need to consider the stratification induced complex computations. The mean values of $\delta^{18}O$ of the lake considered for the mass balance are –8.2‰ (February, 1994) and –7.3‰ (February, 1995), with a net change of 0.9‰. The weighted mean precipitation $\delta^{18}O$ value was –11.3‰, and the mean $\delta^{18}O$ for evaporation using annual average weather conditions was –29.1‰ (most depleted value was –34.2‰). The mean $\delta^{18}O$ values of surface inflow and inflow through the drains were –8.6‰ and –8.0‰, respectively.

The average $\delta^{18}O$ value of the surface outflow has been considered as –8.0‰ as surface outflow occurs mostly during high water levels and during higher surface inflow, with less time for proper mixing. This is shown by the values observed during September 1996, when the surface layers are comparatively depleted than the bottom waters. The mean $\delta^{18}O$ value of groundwater was –9.0‰ and that of the subsurface outflow from the lake has been considered as –8.0‰. Since the evaporation $\delta^{18}O$ was estimated using the Craig and Gordon Model that is very sensitive to the input values, two different sets of $\delta^{18}O$ values have been used to estimate the sub-surface outflow component from the lake. In the first set the $\delta^{18}O$ value of lake evaporate calculated using average weather conditions is used and in the second set the most depleted $\delta^{18}O$ value of lake evaporate was used. The most enriched value (+0.2‰) was calculated for the month of August, when the relative humidity is about 95%. As presented in Section 3.4.6.1 the results of the Craig and Gordon Model are very sensitive to humidity component, and that when relative humidity is above 75%, the results may be unrealistic. Therefore, the Maximum $\delta^{18}O$ value of the lake evaporate has not been considered as representative of annual evaporation isotopic index, in solving the isotopic mass balance equation.

The sub-surface outflow (SSo) from the lake computed using the isotopic mass balance approach has been shown in Table 4.6. The results have been used in equation (4.7) to compute the sub-surface inflow (SSi) to the lake. The results show that the sub-surface components are dominant over other components. The SSi accounts for 51.0% of total inflow and SSo accounts for 58.8% of total outflow, when the results of the isotopic mass balance that uses mean annual

δ_E value are considered. However, the SSi and SSo are higher, 70.2% and 75.0% respectively, when the isotopic mass balance using minimum δ_E is considered. Since, the difference between the results obtained by two different sets of isotopic values are substantial, the question as to which of these two may be a true representation can be addressed by comparing the results with an independent estimate. The results obtained by the conventional water balance compare well with the results obtained by isotopic mass balance that uses mean annual evaporation $\delta^{18}O$ value. The SSo estimated with mean annual δ_E value is about 5% higher than the conventional estimate while that estimated with most depleted δ_E value is 120% higher. Therefore the results obtained with mean annual δ_E value may be taken as closer to the true values of sub-surface components.

4.7.1 Uncertainties in the isotopic mass balance approach:

The uncertainty in the estimation of sub-surface outflow from the lake and in turn in the estimation of sub-surface inflow to the lake can be estimated following the general expression presented by Bevington (1969):

$$\sigma x^2 = \sigma u^2 (\partial x/\partial u)^2 + \sigma v^2 (\partial x/\partial v)^2 + \ldots\ldots\ldots\ldots\ldots$$

Since SSo is a function of all other inflow (except SSi) and outflow components, and also a function of $\delta^{18}O$ values of all the components, the standard error in all the estimates used as inputs to equation (4.6) should be known. The standard errors in the water balance components have been presented in Section 4.5.8. 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 evaporate, can not easily be ascertained.

In the estimation of δ_E (equation 2.16), there are three main variables used viz., δ_L, δ_a and h. The errors in the determination of these variables will be propagated to the estimated isotopic composition of lake evaporate. In the case of a well-mixed lake, the spatial variation of δ_L may not be significant. The error in the computation of δ_a may be as large as 2.5‰ (Kumar and Nachiappan, 1999). However, error in the measurement of third variable h, and in the computation of its mean value is generally considerable [*Linsley et al.*, 1975]. The propagated error or the uncertainty in the estimation of δ_E, due to the errors (σ^2) in the above input variables as suggested by Bevington (1969) can be described as:

$$\sigma\delta_E{}^2 = \sigma\delta_L{}^2 (\partial\delta_E / \partial\delta_L)^2 + \sigma\delta_a{}^2 (\partial\delta_E / \partial\delta_a)^2 + \sigma h^2 (\partial\delta_E / \partial h)^2 \qquad (4.8)$$

From equation (2.16), we get

$$\partial\delta_E / \partial\delta_L = \alpha^* / (1-h) \qquad (4.9)$$

$$\partial\delta_E / \partial\delta_a = -h / (1-h) \qquad (4.10)$$

$$\partial\delta_E / \partial h = (\delta_L - \delta_a - \epsilon^*)/(1-h)^2 \qquad (4.11)$$

Combining equations (4.8) to (4.11),

$$\sigma\delta_E{}^2 = \sigma\delta_L{}^2 (\alpha^* / (1-h))^2 + \sigma\delta_a{}^2 (-h / (1-h))^2 + \sigma h^2 ((\delta_L - \delta_a - \epsilon^*)/(1-h)^2)^2 \quad (4.12)$$

From the first two terms on the right hand side of equation (4.12), it is seen that the propagation of errors in δ_L and δ_a are controlled by the values of humidity. As h approaches unity, the errors tend to infinity. 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 from Figure 4.5. The propagated errors due to errors in input isotopic variables, 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. Since, the uncertainty in the calculated δ_E, when the error in the computation of humidity is considered, is unrealistic (>800%), it has been omitted in the estimation of propagated errors. The estimated error in the calculated δ_E is about 4‰. Using the above values of standard error in the input variables, the error in the estimated SSo has been computed. The results show that the uncertainty in the SSo computed using annual mean δ_E value is about 212% and that in the SSo computed using minimum δ_E is about 99%. Although the estimates of error in the estimated components are too high, in reality

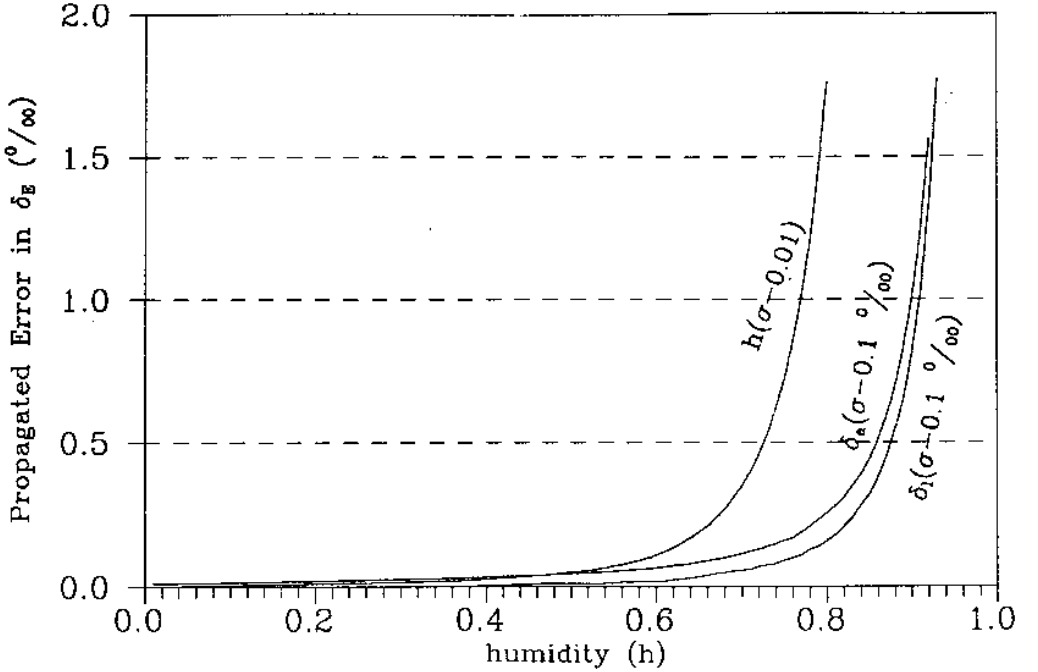


Figure 4.5 Propagated error in δ_E in CGLR model results due to errors in input parameters at different values of h

this may not be so. The higher values are in the uncertainty are indicative of the sensitivity of the approach used.

The method is also highly sensitive to the difference between the $\delta^{18}O$ values of the groundwater inflow and that of the lake seepage. The relative error decreases with increase in the difference between these two isotope indices used. LaBaugh, et al. (1997) in a similar study have adopted an uncertainty of 50% in the subsurface components of the lake – not based on the classical propagated error estimation approach – but based on the errors in an alternate method (flow-net) that they used in addition to isotope mass balance and inorganic chemical mass balance methods. Therefore, in the present investigations also, a similar approach may be adopted and a conservative estimate of 10% may be considered as the uncertainty in the estimations of sub-surface components.

4.8 Chlorine Mass Balance

Mass balances of other conservative chemical constituents provide an additional means by which SSi and SSo can be evaluated. Since chlorine is a conservative chemical, it may be used for the purpose. The advantage of the chlorine mass balance over the isotope mass balance method is that, the mass of chlorine lost from the lake through evaporation is zero. Therefore, it is much simpler, compared to isotope mass balance. The disadvantage is that, chlorine may be introduced into the lake – groundwater system through anthropogenic activities.

The concentration of chlorine in lake water was 8 mg/L and 10mg/L during February 94 and 95 respectively. The mean concentration in the drain water (Di) was 31 mg/L, and that of surface inflow (Si) was 24 mg/L. The latter value was used considering the fact that during monsoon period the flow in the drain waters (particularly during the sampling period) was dominated by channelled surface runoff. The mean concentration of chlorine in groundwaters was 16 mg/L in the upstream springs viz., Pardhadhara, Alma cottage and Lakeview springs. The mean chlorine concentration of the downstream springs Sipahidhara and Gupha Mahadev of 17mg/L has been considered as representative of subsurface outflow from the lake. The input of chlorine by precipitation has been considered as 1 mg/L (cf. Section 3.2.5).

The sub-surface outflow (SSo) from the lake computed using the chlorine mass balance approach has been shown in Table 4.6. The results have been used in equation 4.7 to compute the sub-surface inflow (SSi) to the lake. The results corroborate the findings of isotope mass balance method i.e., is the sub-surface components are dominant over other components. The SSi computed using the chlorine mass balance method accounts for about 55.0% of total inflow and SSo accounts for about 59.0% of total outflow. Compared to the estimates by conventional water balance, the SSo computed by chlorine mass balance method is higher by 26%.

5.0 CONCLUSIONS

The results presented in Table 4.6. show that the estimates of sub-surface inflow to the lake and outflow from the lake obtained through the isotopic and chemical balance compare very well with those obtained through conventional water balance method. The water retention time (τ) of the lake, defined as the ratio of total volume of the lake to the total annual inflow to the lake (V/I), is a better parameter to compare the results as it reflects the results in relation to the lake size. τ computed using isotopic mass balance approach is about 1.93, chlorine mass balance is about 1.77 and conventional water balance is about 1.92.

Table 4.6 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.

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.

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