NEOGLACIAL CLIMATE IN THE SOUTHERN COAST MOUNTAINS, BRITISH COLUMBIA

By

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Abstract

Palaeobotanical records of Holocene climate change in the southern Coast Mountains identify a cooler/wetter Neoglacial period subsequent to 6600 BP. Geomorphic evidence of alpine glacier advance suggests that there were three distinct cooler/wetter periods during the Neoglacial, but this pattern has not been identified in palaeobotanical studies.

By careful selection of a sensitive alpine site this thesis has recognised this structure in a palynological record of Neoglacial climate. This continuous record of Neoglacial climate which has the same basis as records of early Holocene climate (i.e. palynological) and hence allows more direct comparisons of the two periods.

Pollen spectra, conifer needle macrofossils, organic matter content, and magnetic susceptibility were assessed for a 4800 year continuous sequence of sediment from an alpine lake. Calibration of the *Picea/Pinus* pollen ratio by using an altitudinal transect of surface pollen samples allowed partial quantification of shifts in treeline. Treeline at the site was at least 85 m above the present level from 4800-3800 BP, suggesting that summer temperatures were at least 0.6°C above the present. High treeline until 3800 BP indicates a relatively late date for the Hypsithermal/Neoglacial transition at this site. Alternatively, the apparent complexity of this transition in the Coast Mountains may be due to difficulties of separating temperature and precipitation signals in many climatic records. Treeline declined to near present levels by 2500 BP and was lower than present from 2500-1500 BP and from 1200 BP until close to the present.

Estimates of equilibrium line altitude depression for Coast Mountain glaciers during the Little Ice Age suggest that these periods of lower treeline were due to a cooling of up to 0.8°C. During the last 5000 years the Southern Coast Mountains have experienced fluctuations on the order of 1.5°C.

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Acknowledgments

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Chapter I Introduction

Isotope evidence from ocean cores has clearly identified the major glacial/interglacial cycles of the Quaternary Period (Shackleton and Opdyke, 1973). However the relatively low sedimentation rates associated with oceanic sediment preclude analysis of high frequency climatic changes. The need to characterise natural high frequency climatic changes has become particularly acute, because without such information a proper assessment of potential anthropogenic climate change cannot be made. We must therefore turn to the more fragmentary terrestrial record for evidence of the nature of past high frequency climatic change.

The objective of this thesis is to identify, and attempt to quantify the climate changes of the Neoglacial period in the southern Coast Mountains of British Columbia. Palaeobotanical and physical evidence from lake sediments are used to reconstruct the environmental history of the alpine zone. Ratios of lowland *Pinus* (Diploxylon) to subalpine *Picea* pollen types for the study area are used to provide a sensitive, and quantifiable record of changing treeline, and hence climatic change. Supporting evidence is derived from estimates of the equilibrium line altitude (ELA) depression of Little Ice Age glaciers.

1.1 Holocene Palaeoclimate in Southern British Columbia

Evidence of Holocene palaeoclimate in southern British Columbia is derived almost exclusively from two sources. Palaeobotanical studies have elucidated the patterns of vegetational change in the region. Many of these changes have been interpreted as climatically forced, thus providing evidence of past climatic change. Geological evidence of past glacier fluctuations provides the second line of evidence. The nature of the palaeoclimate record provided by these two lines of evidence in Southern British Columbia is considered below.

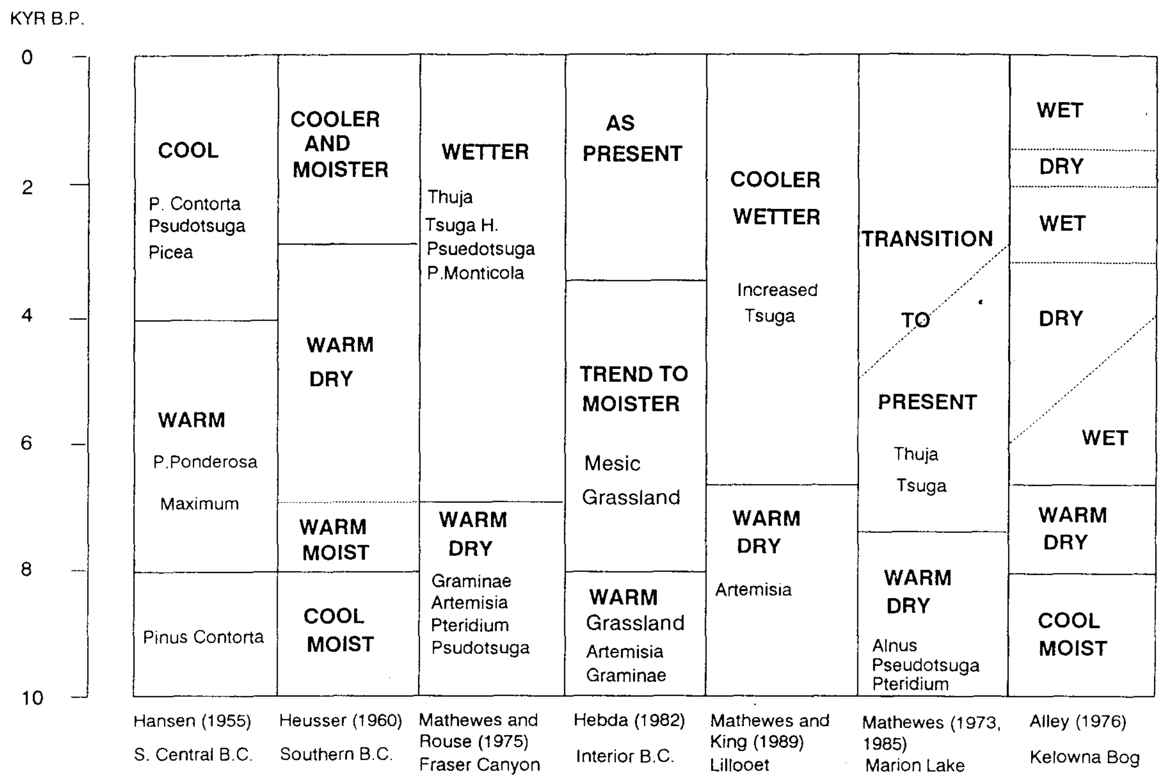
1.2 Palynological Studies in Southern British Columbia

The earliest palynological studies of southern B.C. were conducted by Hansen (1940, 1947, 1950, 1955), but interpretation of his pollen diagrams is hampered by the lack of firmly established dates for the sedimentary sequences from which the pollen was extracted. However these early studies did succeed in identifying the broad patterns of Holocene vegetational and climatic change in the region. Hansen emphasised successional processes when explaining the changing relative abundance of species, however he also conceded a role for climatic change, postulating a shift from warm, dry early Holocene climates to cooler, wetter late Holocene conditions.

Subsequent pollen diagrams associated with well dated sediment sequences show a similar trend. However as Mathewes and Heusser (1981) noted, the early Holocene warm period or Hypsithermal has been shown to be a time transgressive phenomena in contrast to the original definition of the Hypsithermal as a time stratigraphic unit (Deevey and Flint, 1957). A summary of climatic trends derived from palynological work in southern B.C. is presented in Figure 1.1. An index map of sites referred to in this review is presented in appendix 1.

Some broad similarities emerge from these studies. All conform to the general trend of cooling, and increasing moisture proposed by Hansen. The work of Alley (1976) is the one exception. This study based on bog sediments from the dry interior shows alternating wet and dry periods throughout the Holocene. In the other diagrams the timing of the onset of late Holocene cooling varies. The work of Hansen, 1955; Heusser, 1960; and Hebda, 1982, suggests that conditions warmer than present

Figure 1.1 Summary Diagram of Holocene Climatic Trends in Pollen Diagrams from Southern British Columbia.



First two columns on the left after Alley (1976) remained until 3500-4000 B.P. In contrast the sites studied by Mathewes and colleagues indicate a much earlier onset of cooling around the time of the Mazama ash fall dated at 6840 BP (Sarna-Wojcicki *et al.*, 1983).

In addition to the qualitative interpretation of pollen data, work by Heusser has initiated a more quantitative approach to palynology in the region. Heusser *et al.* (1980) collected surface pollen data from 178 sites between San Francisco and Alaska. Based on this data and available climate information, a transfer function was developed to relate pollen data to mean July temperature and annual precipitation. Mathewes and Heusser (1981) applied this transfer function to the pollen data from Marion lake, U.B.C. Research Forest, to produce temperature and precipitation plots for the last 12 ka. This quantitative treatment of the Marion Lake data illuminates essentially the same pattern as the earlier qualitative analysis of Mathewes (1973). A warm dry period lasting until after 8000 BP was succeeded by cooler wetter conditions, especially subsequent to 6840 BP. The principal contributions of the quantitative analysis are the numerical estimates of the range of temperatures and rainfall experienced during the Holocene. Whilst the estimates of past conditions may not be absolutely correct, the magnitudes of the relative changes may be instructive. The study suggests that at the height of the Hypsithermal, mean July temperatures may have been as much as two degrees Celsius above the present. Annual precipitation was up to 800 $mm.yr^{-1}$ less than modern values.

In a detailed review of the Holocene climates of southern B.C., Mathewes (1985) concluded that the early Holocene warm period peaked at around 7500 BP. Subsequent to the Mazama ash fall at 6840 BP there was a "poorly defined period of climatic transition" (p.419) which entailed a gradual cooling, and an increase in precipitation until modern conditions were established around 4500-3000 BP.

The nature of the transition from warm, dry Hypsithermal conditions to cooler, wetter late Holocene climates is one of the most problematic issues raised by palaeoclimatological work in the region. Some sites indicate relatively abrupt cooling after 6840 BP. In particular, the quantitative reconstruction from the Marion lake core (Mathewes and Heusser, 1981) suggests that temperatures as cool as, or even slightly cooler than the present, were established by that time. In contrast, Hebda (1982) suggests that in the interior conditions remained warmer than at present until around 4000 BP. Mathewes (1985) suggestion that modern conditions were not established until 4500-3000 BP carries a similar implication. Clearly as Mathewes suggested this was a period of complex transition. No doubt there was spatial variation in the nature of the climatic changes occurring between 7000 and 4000 BP in southern B.C. However the evidence to date does not present a clear spatial pattern. Given the point nature of sites for palaeoclimatic reconstructions, and the highly variable topography and hence local climates of the region, such variability is perhaps to be expected. More rigorous consideration of the sources of local variability may be necessary to clearly identify the nature of the transition in regional climate.

1.3 Geomorphological Evidence of Holocene Glacier Fluctuation

In addition to palynological evidence, the second major source of palaeoclimate data for southern B.C. comes from evidence of periods of glacier advancement, presumably associated with climatic deterioration. As indicated by the discussion of palynological investigation above, the early Holocene was marked by warm, dry conditions. These conditions are unfavourable to glacier expansion. The principal periods of Holocene glacial advance are concentrated in the cooler, wetter second half of the Holocene. Mathes (1939) used the term "Little Ice Age" to describe the regeneration of glaciers in western North America. By reference to the work of Antevs (1938) and Gales (1915) on limnological changes in the Sierra Nevada, he inferred a date of 4000 BP for the start of this period (Porter and Denton, 1967). The term Little Ice Age has subsequently come to be used predominantly to describe the glacial advances of the last 500 years. Porter and Denton (1967) advocated the use of the term "Neoglacial" to describe glacial advance subsequent to the Hypsithermal interval. By definition therefore, the Neoglacial interval is a time transgressive phenomenon.

Evidence of former glacial advances may be derived from terminal moraine positions, till sheets, trim lines and sequences of lateral moraines. Such features may provide good evidence of the former extent and thickness of the glacier. However, at a single site, a complete record of fluctuation may not be available since extensive later advances may overrun and rework evidence of less extensive earlier advances. It is therefore necessary to develop a regional chronology of glacial advances from a variety of well dated sites. The principal techniques which have been utilised are listed in Figure 1.2. Palaeoclimatic interpretation of glacier expansion chronologies rest on the assumption that glaciers respond to climate change. Porter (1977) demonstrated that variation in ablation season temperatures and accumulation season precipitation explain most of the temporal variation of glacier equilibrium line altitudes in the Olympic Mountains. However, Bradley (1985) noted that climatic interpretation of glacier advance may be complicated by local influences on glacier mass balance such as the following:

1) For some glaciers, particularly small cirque glaciers blown snow or snow avalanched onto the glacier surface may be important components of the mass balance.

2) Some glaciers respond to a negative change in mass balance by downwasting rather than frontal retreat.

3) Glaciers exhibit differing lag times in their response to climatic change, largely dependent upon there size. They therefore respond to different frequencies of climatic change. (This is less of a concern for studies spanning millennia).

Figure 1.2 Dating Techniques Applicable to Moraines and Till Sheets in the Southern Coast Mountains, British Columbia

Technique	**Results**	**Example**
Dendrochronology	Relatively thin rings date glacier approach. Ring counts give minimum age for newly colonised features.	Mathews (1951)
Lichenometry	Minimum dates for ice retreat.	Luckman and Osborne (1979)
Tephrachronology	Tephra above/below drift gives minimum/ maximum dates.	"
Radiocarbon dates	Logs - maximum/minimum dates dependent upon the woods history.	Ryder and Thompson (1986)
	Stumps - date glacier advancing.	"
	Peat/Soil - above/below gives minimum/maximum date.	"
Historical Records	Records of past terminus positions, or phases of advance.	Luckman (1986)
Stacked Till Sequences	Relative dating	Ryder and Thompson (1986)

4) The behaviour of surging or calving glaciers is largely dependent on internal system dynamics and therefore not interpretable in terms of climate. (Sugden and John, 1976; Mann, 1986)

Confusion may also arise at sites exhibiting "pseudo-moraines" these features are produced by rockslide onto a glacier. This reduces summer ablation at the glacier surface causing an advance. When the glacier retreats the slide debris is deposited as a moraine-like feature (Ryder, pers. comm.). Such glacier advances, and the associated 'pseudo-moraines' are not directly associated with climatic change. Lack of evidence of a specific advance at a given site may therefore be ascribed either to particular local conditions, or to destruction of evidence of the event by later advances. Given these sources of uncertainty within the record, the importance of building a regionally consistent chronology based on multiple sites is further emphasised.

The earliest study of Neoglaciation in the southern Coast Mountains of B.C. was conducted by Mathews (1951). Late Neoglacial moraines in the Garibaldi map area were dated by dendrochronology to the early 18th and mid-19th centuries. Based on the lack of evidence for other post-Wisconsinian advances, Mathews suggested that the late Neoglacial advance was the most extensive of the Holocene glacial phases. Subsequently, Stuiver *et al.* (1960) radiocarbon dated overridden stumps in the area to 5300 $\pm$ 200 BP and Lowden and Blake (1975) also dated stumps to 5300 $\pm$ 70 BP suggesting an earlier phase of glacier advance.

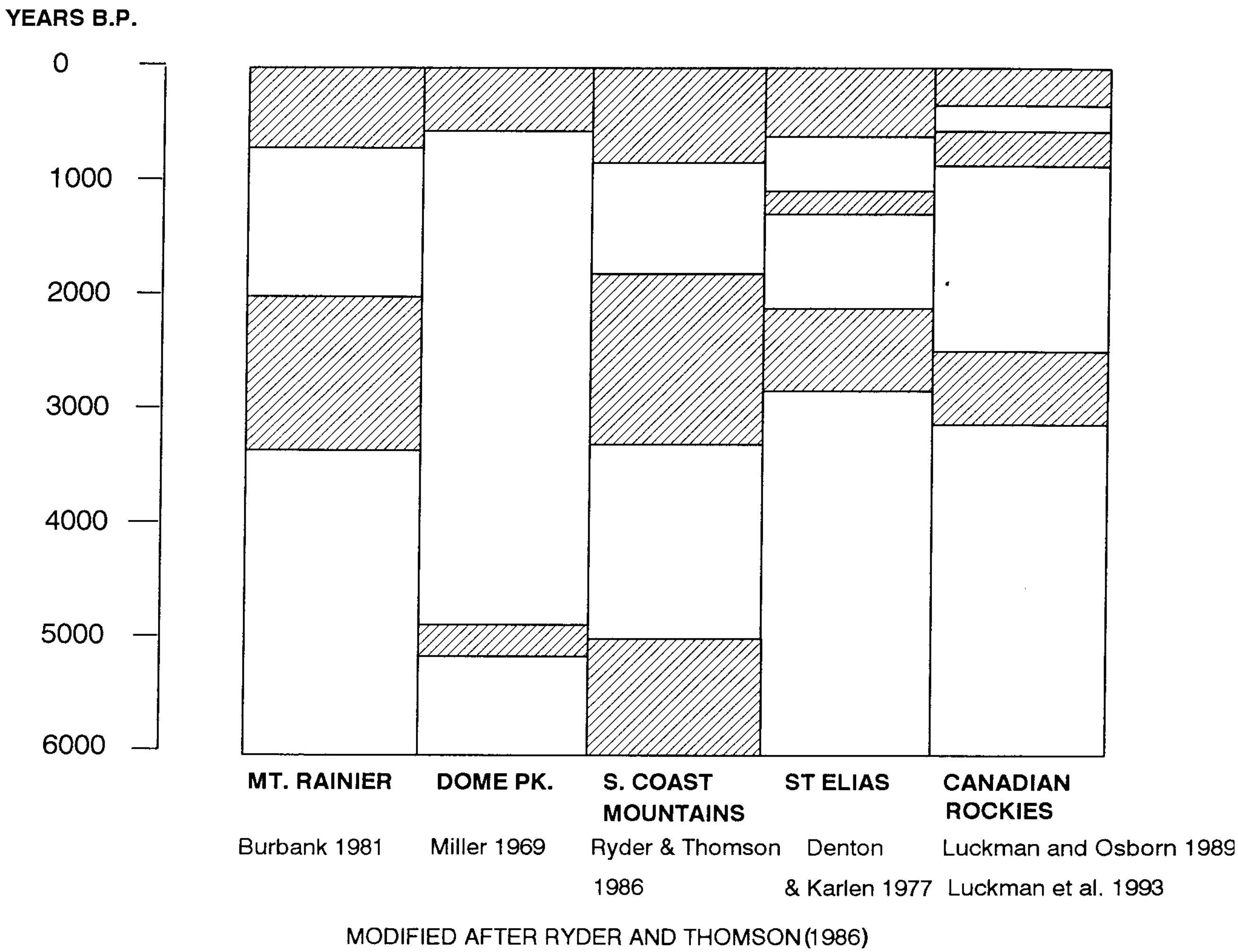
The most comprehensive study of Neoglaciation in the southern Coast Mountains was conducted by Ryder and Thomson (1986). They present radiocarbon dates and interpretations of moraine morphology and stratigraphy for five glaciers. Interfingered till and fluviatile sedimentary units from lateral moraines were interpreted as representing different phases of advance. Dating of incorporated organics and palaeosols allowed the development of chronologies at single sites even where terminal moraine evidence was lacking. This study named the mid-Holocene advance proposed by Mathews the "Garibaldi Phase". Two further periods of advance were identified. These were the "Tiedemann phase" 3300-1900 BP so called because it appears to have been particularly extensive around the Tiedemann Glacier, and the Late Neoglacial advance dating from 900 BP to around 100 BP. Studies of lake sedimentation rates in a glacially fed catchment (Souch, 1990) support this tripartite division of the Neoglacial interval.

A large body of literature also exists on Holocene glacial fluctuations in mountain ranges adjacent to the Coast Mountains. The periods of advance identified by these studies as well as those recognised by Ryder and Thomson (1986) are summarised in Figure 1.3. It is apparent that the mid and late Holocene advances are widely recorded. In contrast, evidence for an early Neoglacial advance synchronous with the Garibaldi Phase is less extensive. The temporal correlation of periods of glacier advance across the extended region allows some confidence in the fact that the observed glacier fluctuations are climatically forced. Denton and Karlen (1973) identify a 2.5 kyr periodicity in northern hemisphere glacial advances during the last 15 kyr. The last three phases of advance approximate the phases identified above. It may therefore be that the glaciers of the northwestern Cordillera have responded to the regional manifestations of hemispheric climate changes.

1.4 Neoglacial Climate in the Coast Mountains of B.C.

As noted above, the Neoglacial is a time transgressive phenomenon. In the Coast Mountains the evidence of Ryder and Thomson (1986) suggests that it corresponds to the period subsequent to the Mazama ash fall. There is general agreement that the Neoglacial marks a cooler and wetter phase than the preceding Hypsithermal, however

Figure 1.3 Neoglacial Advances in Southern British Columbia and Adjacent Areas



the evidence regarding the detailed nature of Neoglacial climate in the region is somewhat contradictory. The geomorphological evidence of glacier fluctuations suggests that climate must have become cooler and or wetter at three distinct times during the Neoglacial. Some amelioration would be expected between these cooler phases. In contrast, the climatic signal interpreted from pollen evidence exhibits much less structure. Pollen evidence suggests either gradually declining temperatures until 4000 BP or else a rapid decline to temperatures as cool or cooler than the present by 6840 BP. In either case the (presumably) climatic events associated with the phases of glacier expansion are not recorded. Only the pollen diagram from Kelowna Bog (Alley, 1976) exhibits structure within the Neoglacial interval. The wetter phases identified by Alley are not closely dated, however he tentatively correlates them with three Neoglacial stades.

Parallel lines of evidence support the tripartite division of the Neoglacial apparent in the glacial record. For example, Jones (1988) identified three phases of reactivation of earth flow complexes in southern B.C. Mathewes and King (1989) similarly identify periods of higher lake levels near Lillooet which corrrespond to the phases of glacial advance. Such evidence supports the contention that Neoglacial climate change has been more complex than the pollen record suggests.

1.5 Research Question

The strongest evidence of structure in the Neoglacial climate record of the Coast Mountains comes from the glacial record. Unfortunately this record is fragmentary, and discontinuous in time and space. Further, interpretation of palaeoclimate from glacier fluctuations is complex, since both temperature and precipitation affect the mass balance. Quantitative interpretations are especially difficult. It would therefore be preferable to have evidence of the detailed nature of Neoglacial climate from the pollen record. This has several practical advantages; pollen is easily extracted from lake and bog sediments which are widespread and usually undisturbed. Sedimentation at such sites is normally continuous. The difficulties of destruction of evidence by later events are therefore avoided. In addition ecological interpretation of pollen stratigraphy may allow estimates of relative changes in warmth or wetness.

Souch (1990) noted that evidence of early Holocene climate in southern British Columbia has primarily been derived from palynology. In contrast evidence of Neoglacial climate is primarily geomorphological. She suggests that the differing climatic responses, lag times, and sensitivities of these two lines of evidence make comparisons difficult. Detailed palynological evidence of Neoglacial climate, directly comparable to the palynological record of early Holocene climate, would aid a more complete understanding of the nature of Holocene climatic change in the region.

The aim of this study therefore, is to elucidate the detailed nature of Neoglacial climate change in the Coast Mountains based on a continuous sediment record. If the climatic changes associated with Neoglacial advances can be identified in the pollen record, a more detailed picture of the nature of these changes and the intervening climatic regime should emerge. In addition, such a record will provide further evidence as to the timing of the onset of the Neoglacial in the region.

Chapter two develops an approach to the palynological study of Neoglacial climate in the region. In Chapters three to seven this approach is then developed and applied to a site in the southern Coast Mountains. In order to develop as complete as possible picture of Neoglacial climate change it is necessary to examine multiple lines of evidence. Chapter five therefore considers other proxy data derived from a continuous sedimentary record. Similarly Chapter eight examines the extent to which the geomorphological record of glacier advance can provide complementary evidence of Neoglacial climate change. Chapter eight is based on evidence from glaciers across a transect of the Coast Mountains from the coast to the interior. It therefore provides a regional perspective which complements the site specific data derived in Chapters three to seven.

Chapter II Approach

2.1 Identifying "Sensitive" Pollen records

The most widely used technique of climatic reconstruction applied to terrestrial sedimentary environments is pollen analysis. As noted previously, pollen studies from southern British Columbia have not clearly identified the Neoglacial climatic fluctuations of interest. Ryder and Thomson (1986) suggest that most glaciers were at their greatest Neoglacial extent during the Little Ice Age (LIA). Presumably this was the period of maximum climatic deterioration. The LIA is a well documented global cooling event during which summer mean temperatures were depressed by 1 - 2°C (Grove, 1988). Detection of such minor climatic fluctuations in the pollen record requires that small changes in climatic parameters produce relatively large changes in vegetation composition at the study site.

In coastal B.C. the natural dominant vegetation is coniferous rainforest. Lowland sites are wet and relatively warm. The major coniferous species are therefore growing in conditions well within their climatic tolerance, and are relatively insensitive to minor climatic fluctuations. Most of the pollen analytical work done in southern B.C. has been from low elevation sites. It is therefore not unexpected that minor Neoglacial fluctuations in climate are not registered (e.g. Mathewes and Heusser, 1981). In order that minor fluctuations in climate produce major shifts in local vegetation it is necessary that the site under investigation be close to a major climatically controlled ecotone.

In a given climatic zone the steepest climatic gradients are typically associated with elevational change. The most marked elevationally defined ecotone in the Coast

Mountains is treeline[1]. It was therefore anticipated that a study of pollen from a continuous sediment sequence close to treeline will identify a vegetational response to Neoglacial climatic fluctuations. Supporting evidence for this supposition is found in a rare high altitude pollen study from British Columbia. Jones (1987) produced a pollen diagram from Red Mountain in south central B.C. at an altitude of around 6000 ft. close to treeline. The key feature of this diagram is a tripartite division of the *Alnus* curve. The curve describes a two stage decline in *Alnus* values, which chronologically approximates to a decline in *Alnus* during the mid and late Neoglacial periods. The interpretation made from this pattern was of progressive thermal deterioration during the Neoglacial, which was particularly marked during the phases of glacier expansion.

This thesis therefore hypothesises that if a suitable continuous sedimentary sequence is found at treeline in the Coast Mountains it will be possible to identify in the pollen record vegetational responses to Neoglacial cooling. Such information may cast new light on the nature of Neoglacial climatic change.

2.2 Quantification

The treeline ecotone has been widely studied. In particular, Tranquilini (1979) clearly outlined the effects of severe alpine climate on tree physiology. Tranquilini notes that since Humboldt first postulated thermal control of treeline it has been recognised that the altitudinal limit of tree growth coincides with the level of the ten degree July isotherm. Tranquilini suggests that in fact the ultimate control on treeline altitude is winter desiccation. However, the mechanism by which this is induced is that there is an inadequate period of warmth for new shoots to fully develop and become resistant to

1 In this thesis the term treeline will be used to refer to the limit of upright growth forms of height less than around 2 m. This is a somewhat arbitrary, and rather anthropocentric definition of what is a normal or stunted growth form. But, at least at the study site this is approximately the size at which trees begin to exhibit the unusual growth forms associated with krumholz. The commonly used term timberline is a commercial term and is avoided.

desiccation. It is therefore likely that the 10 degree isotherm is a proxy for a certain number of degree days of warmth. Temperature therefore is a critical determinant of treeline altitude.

It is likely that temperature sets a theoretical maximum altitude but actual altitude may be lower due to other environmental constraints. Too much precipitation falling as snow may lower treeline as summer seedling establishment may be delayed where snow lies late into the summer. In glaciated areas treeline is often locally depressed by edaphic factors. Where Little Ice Age glacial advances have left large areas of fresh moraine insufficient time may have elapsed for colonisation and growth of trees. Further, soil development has often been insufficient to support trees. The effect of insect infestation is highly variable in space and time but may act to locally lower treeline. Despite this apparent complexity, it should be possible, through careful site selection, to control for such factors as the above, so that treeline altitude is a function of local thermal regime.

The relation between temperature change and elevation is described by the mean mountain lapse rate for the area (Barry, 1992). A further advantage therefore, of studying the treeline ecotone is that changes in treeline elevation may be interpreted directly as representative of change in the thermal climate. If it is possible to quantify the vertical extent of treeline advance or retreat it will be possible to produce a quantitative record of Neoglacial thermal climate.

Semi-quantitative evidence of former treelines altitudes has been derived from radiocarbon dating of dead snags above current treeline (e.g. Lamarche, 1973; Denton and Karlen, 1976; Luckman *et al.*, 1993). This provides a minimum estimate of former treeline altitudes. However due to the vagaries of preservation nothing may be interpreted from the absence of dead wood. In addition, no record of treeline below the present level is preserved. Fossil wood therefore provides a temporally discontinuous record of higher treelines. Macrofossil evidence (e.g. wood, needles, seeds, etc.) from bogs and lakes provides strong evidence of local vegetation. Therefore, for sedimentary basins at treeline, where the ecotone may be migrating back and forth across the site, a record of changing treeline may be preserved. However such a record is primarily of presence or absence of trees from the site and does not permit quantitative estimates of treeline shifts. Records from an elevational series of lakes/bogs would provide a quantifiable, but discontinuous record.

What is required to obtain a continuous quantifiable record of treeline altitude is a proxy record which registers the position of treeline even when it is remote from the site. Pollen provides a possible source of such information. Pollen is well dispersed such that pollen rain consists of local, extra local, and regional components (Jacobsen and Bradshaw, 1981). Change in the local and extra local component should reflect whether the site is above or below treeline at a given time. In addition, relative change in the extra local component might reflect a quantitative estimate how far above or below treeline the site is, given that pollen transport is subject to log linear distance decay, (Colwell, 1951 (cited in Maher, 1963); Tauber, 1965). This reasoning has been the basis of a variety of attempts to provide a quantitative record of treeline shifts based on the pollen record (e.g. Maher, 1972; Beaudoin, 1986). These studies utilised techniques developed by Maher (1963) (see 2.3.3). This study will adopt the approach to quantification of the pollen record developed by Maher in an attempt to produce a record of Neoglacial treeline variation in the southern Coast Mountains.

Interpretation of pollen diagrams from topographically complex areas is difficult (Faegri and Iverson, 1990). Uncritical application of simple techniques is therefore fraught with dangers of misinterpretation. A conceptual model of local pollen deposition is vital to correct qualitative or quantitative interpretation of pollen data. A detailed review of the literature on pollen deposition at treeline and interpretation of treeline pollen records was therefore undertaken. This review identifies the theoretical problems and possibilities of high altitude palynology, in an attempt to establish whether a high altitude pollen record might reasonably be expected to register Neoglacial climate fluctuations. The assumptions of Maher's (1963, 1972) approach to quantifying the record are examined in order to provide a theoretical framework for the interpretation of any such record.

2.3 Review of Literature on High Altitude Palynology

2.3.1 Altitudinal Zonation of Modern Pollen Deposition

Studies of modern pollen deposition in mountainous areas are relatively rare. Analysis by Heusser (1973) of 25 moss polster samples from varying altitudes on Mt. Rainier, Washington State, showed clear differentiation of the alpine zone from lower elevations based on *Alnus* pollen, which dominates the lower zones. The most important alpine pollen types are *Cyperaceae* and *Tubuliflorae*. The importance of the pollen of alpine indicator species for differentiating alpine and subalpine sites was emphasised by Mathewes (1988). He demonstrated that the surface pollen spectra from Swiss alpine lake sediments could be clearly distinguished those from sediments of lakes in the montane forest. Subjective assessments were confirmed by conducting a principal components analysis of the data. Jacobsen and Bradshaw (1981) demonstrated theoretically that with decreasing size of sedimentary basin, representation of local as opposed to regional pollen in the sediments increases. Mathewes (1988) demonstrated that in the Swiss alpine zone, high representation of alpine herbs is only found in a very small pond (0.002 ha.). This finding has important methodological implications, especially in the light of the suggestion by Solomon and Silkworth (1986) that only identifiably local pollen types provide valid information on vegetation distribution in mountainous areas. They measured altitudinal variation in pollen deposition in the southern Sierra Nevada. Their study was based on pollen trap data collected at 23 sites over a period of six years. Pollen deposition was shown to be strongly associated with local vegetation types. Pollen deposition by a given species was shown to decline by a factor of 5-10 between locations where the species is present in local vegetation, and those where it is not. The study therefore concluded that there is a "step change" between high pollen deposition values at the site of the vegetation, and relatively uniform low values beyond the boundary of the particular vegetation site.

The suggestion that altitudinal zonation of pollen deposition is characterised by a step function has important implications for attempts to reconstruct changes in altitudinal vegetation zonation from the pollen record. The reconstruction of shifts in vegetation away from a particular site based on pollen accumulation at that site rests on the assumption that the distance decay in pollen deposition away from the source plant is described by some continuous function. If, as Solomon and Silkworth suggest, there is a step change between high and low values, then the value of pollen evidence of past vegetation at a site is limited. Only presence/absence of a particular vegetation type at the site based on local pollen types may be inferred.

The examples cited above suggest that a well defined relation exists between altitudinal zonation of vegetation and surface pollen samples along an altitudinal transect. However, other studies of surface pollen samples in mountainous areas have arrived at quite different conclusions. Maher (1963) collected 35 samples of lake sediment and moss between 6850 and 12600 feet in the San Juan Mountains of Colorado. When the pollen percentage data were plotted against altitude there was no readily apparent trend. Kearney (1983) was able to distinguish montane forest surface sediment samples from those collected in sub-alpine forest, in Jasper National Park, based on changing patterns of conifer pollen percentage values. The pollen percentages for these zones were closely linked to the canopy dominance of the relevant species. However Kearney's percentage data showed very poor differentiation between subalpine forest and alpine tundra sites. It appears therefore, that considerable variation exists with respect to the extent to which surface pollen spectra represent altitudinal zonation of vegetation.

Some of this apparent variation may stem from the different measurement techniques used by different authors and varying methods of presenting the data. For example, Solomon and Silkworth's (1986) data are mean annual deposition rates for particular pollen types derived from traps. Pollen deposition rates are extremely difficult to derive from surface lake sediment samples and moss polsters, since the time period of deposition is not known. Therefore Heusser's (1973) and Mathewes' (1988) data are expressed in percentage terms. Pollen representation in percentage based surface spectra is affected not only by pollen production and dispersal, but by relative differences in production and dispersal between species.

Fall (1992) has considered the influence of the various techniques of sampling modern pollen on surface pollen spectra. By comparing pollen counts from surface lake sediments, bog surfaces, moss polsters, and pollen traps, along an altitudinal transect; it was determined that the best representation of local vegetation came from traps and moss polsters. It is however unclear from the analysis whether poorer representation in lake sediments is due to the nature of pollen recruitment and preservation in lake sediments, or simply a function of the size of the source area sampled by the various methods. Jacobsen and Bradshaw (1981) emphasise the importance of lake basin size in determining the source areas of pollen deposited in lakes. In large lakes locally distributed pollen may not reach the centre of the lake whereas it may more easily reach moss polsters close by. In Fall's study, lakes provided the pollen samples least representative of local vegetation, whilst small bogs trapped pollen in a manner similar to moss polsters. Despite this, even in lake sediment samples, subalpine sites could be distinguished from alpine sites.

Of the examples considered above, Mathewes (1988), Maher (1963), and Kearney (1983), studied surface lake sediments and arrived at differing conclusions.

The causes of regional variability in altitudinal zonation of modern pollen deposition and its relation to local vegetation therefore remain unresolved. Reasons for the apparent variability are considered below.

2.3.2 Possible Controls on the Altitudinal Zonation of Modern Pollen Deposition

Kearney (1983) suggested that poor differentiation of vegetation zones by pollen data can be accounted for by "swamping" of the alpine meadow/tundra signal by long distance transport of lowland pollen. Two phenomena combine to produce this effect.

1) Markgraf (1980) collected pollen from traps along an altitudinal transect on the Niederhorn, Switzerland. Annual pollen production at alpine sites was shown to be considerably less than at lower elevations with arboreal vegetation. Faegri and Iversen (1989) state that as a general rule higher altitude plant communities produce less pollen. Therefore even if only a relatively small proportion of lowland pollen is well dispersed to higher altitudes it may significantly distort the representativeness of alpine surface pollen spectra. By comparing monthly pollen trap data with meteorological data, Markgraf was able to characterise, and attempt to account for, the nature of pollen transport between altitudinal zones.

2) Mountain regions characteristically have topographically controlled wind systems. These consist of anabatic (up-valley) winds during the day, and katabatic (downvalley) winds at night. Markgrafs' (1980) data showed that only the anabatic winds are significant in transporting pollen, suggesting that by dawn when the strongest katabatic winds develop most of the pollen released during the day has settled out (p.144).

The "swamping" that Kearney identifies is therefore a product of very low pollen production in the alpine zone, and upslope transport of pollen from high pollen producing lowland areas. The operation of this dual mechanism is well illustrated by pollen data for Rumanian mountains, produced by Frenzel (1969, cited in Faegri and Iversen, 1989). These data suggest that although there is no clear differentiation between surface pollen spectra from the subalpine zone, and those from the upper alpine zone, pollen spectra of the lower alpine zone may be distinct. In the high alpine local production of non-arboreal pollen is low. Because of upslope transport of arboreal pollen, arboreal pollen/non arboreal pollen (AP/NAP) ratios in the high alpine are similar to those from the subalpine forest. However the low alpine meadows are highly productive, especially in the European Alps where treeline has been artificially lowered by clearance for pasture. Therefore pollen spectra of the low alpine may be identified by locally elevated AP/NAP ratios.

Although Markgraf's (1980) data demonstrate the operation of the mechanisms by which "swamping" of alpine pollen production occurs, Markgraf also suggests that there is an altitudinal limit to upslope transport. Topographically controlled wind systems dominate to 1200 m in the Niederhorn. Above this altitude (approximately the elevation of the valley sides) prevailing regional winds are dominant. Therefore surface pollen spectra above 1200 m altitude are more influenced by the regional background pollen rain, than pollen transport from immediately downslope. Solomon and Silkworth (1986) similarly suggest that there is an altitudinal limit to upslope transport. Although the step change in pollen deposition which they identify occurs between upland and lowland vegetation assembleages they do not suggest that a simple change in surface vegetation could produce such a marked change in pollen deposition. Rather, in order to explain the step change, and the lack of pollen mixing, this implies they invoke a physical mechanism. They suggest that the step change in pollen representation demonstrated by their trap data is due to the development of valley inversions. The accumulation of stable cold air in valley bottoms effectively prevents mixing with warmer upper air. Upslope transport of lowland pollen therefore doesn't occur above the inversion 'lid'. Presumably what Solomon and Silkworth identify is a statistical effect. Inversion effects are temporally discontinuous, but would reduce the cumulative annual upslope transport of pollen.

A further potential reason for variable representation of alpine species in surface pollen spectra above treeline may be derived from the observations of Fall (1992). It appears that only 1/5 of the pollen in alpine lakes is from direct atmospheric deposition. The rest is transported to the lake by slope wash and flow in rills and gullies. As noted earlier, alpine vegetation is highly spatially variable, often controlled by microscale moisture differences. Therefore pollen concentrated by washing into lakes may only poorly represent the full mosaic of local vegetation types. The observation that direct atmospheric inputs comprise a small part of pollen deposition in alpine lakes does not however invalidate atmospheric mechanisms producing variability in surface spectra. Presumably much of the pollen reaching the lake by slope wash is originally derived from atmospheric deposition on the slopes.

It is clear from the above that the altitudinal zonation of pollen at a site is closely linked to local meteorological conditions. Complex topography in mountain areas produces extreme variability in meteorological conditions, such that single site measurements are rarely regionally representative (Barry, 1992). It follows therefore that no single model of pollen dispersal in mountain areas is applicable. Therefore in order to properly interpret pollen diagrams from mountain areas local information about the nature of pollen dispersal, derived from altitudinal transects of surface samples is necessary.

2.3.3 Pollen Ratio Studies

Given the inherent variability of the relation between raw pollen data and altitudinal vegetation zonation some method is required to transform such data in order to maximise the signal to noise ratio. Maher (1963) developed the technique of examining ratios between particular pollen types relative to elevation. The problems of dependency associated with percentage analyses are therefore avoided. In order to utilise the ability of pollen data to record vegetational change remote from the site it is necessary to consider both the regional and local pollen signal. Ratios of regional to local pollen types are therefore a potentially sensitive indicator of vegetational change at, and remote from, the site. Maher considered ratios between *Pinus, Picea, Artemisia,* and *Quercus*. In the San Juan Mountains *Pinus* and *Quercus* are montane species, *Picea* is found in the subalpine, and *Artemisia* is found at all elevations. The log of the pollen ratios was used to predict elevation using linear regression. The best predictions were achieved from *Picea:Pinus* ratios ($r=0.91$). Maher showed that *Picea* pollen is considerably less well dispersed than *Pinus* pollen. *Picea* is confined to the subalpine zone. *Picea* pollen is therefore a local indicator of subalpine forest/treeline. *Pinus* pollen in contrast is extremely well dispersed such that deposition is essentially regionally uniform. Since *Picea* pollen deposition peaks in the subalpine zone, *Picea:Pinus* ratios are at a maximum at treeline and decline both above, and below. Therefore, separate regression lines are plotted for surface samples from above and below treeline. The line representing sites above treeline has a negative slope and the below treeline relation having positive slope.

If it is possible to predict the elevation of surface samples from pollen ratios; then, adopting the ergodic assumption, downcore variation in pollen ratios from a single site should provide a record of the changing "apparent elevation" (Maher 1963) of the site with respect to treeline. Since *Picea* is confined to the subalpine this apparent elevation is with respect to treeline. Changing pollen ratios are therefore a proxy for changing treeline elevation. Higher apparent elevations imply lower treeline and *vice versa*. Maher (1972) derived a second *Picea:Pinus* ratio/elevation calibration curve for a site at Redrock Lake, in the Frontal Range, Colorado. He applied this curve to downcore variations in fossil pollen at Redrock lake to reconstruct past apparent lake elevations. The two calibration lines representing sites above and below treeline present a problem of interpretation. Without further independent evidence it is not possible to distinguish which of the apparent elevation curves reflects reality. At Redrock lake, Maher had an independent line of evidence supplied by macrofossils of *Nuphar.* Where these were present, a more clement climate was assumed implying higher treeline. Therefore Maher applied the curve yielding lower apparent elevation.

Macrofossil evidence is extremely well suited as a means to decide which calibration curve is most applicable. In the Redrock lake study (Maher, 1972), macrofossils merely allowed inference of which apparent elevation curve was correct. However the strength of macrofossil evidence is the ability to determine with some certainty the nature of past local vegetation cover. Andrews (1975) used conifer needle macrofossils to determine whether a site was above or below treeline and hence the correct *Picea/Pinus* ratio to use. Obviously, absence of macrofossils does not guarantee that trees were absent from the site. Therefore there is less certainty about past apparent elevations where macrofossil evidence is lacking. However the combination of macrofossils of tree species, and pollen ratio evidence from the same site provide strong evidence of the direction and extent of treeline altitudinal change.

Careful site selection can further strengthen assumptions about the direction of change. For example Kearney and Luckman (1983) produced a curve representing treeline fluctuation in Jasper National Park based on *Abies:Pinus* ratios. The site was 'well above the general treeline' (p.263) such that all the ratios were less than the maximum values recorded at treeline. The assumption made was that only the calibration curve based on above treeline samples was applicable. High treeline at periods indicated by the ratio analysis were confirmed by the presence of *Abies* needles.

2.3.4 Limitations of the Pollen Ratio Approach

Maher (1963) identified three main assumptions on which the application of the pollen ratio technique rests.

1. - The Ergodic Assumption
2. - Uniform Density of Vegetation
   - Stable Vegetation
3. - Size of Lakes for Surface Sample Collection

The ergodic assumption in this case is that the nature of change observed in space between surface samples is exactly equivalent to changes in time between successive fossil pollen assembleages.

The second assumption is based on Maher's finding that variations in vegetation density affected pollen ratios. Therefore reconstructions of treeline based on pollen ratios make the assumption that vegetation density in the area has not changed in time. Obviously, above treeline substantial vegetational change occurs. The requirement is that the density of the lowland forests be relatively constant. This requires that vegetation is in a stable state, not subject to successional change or in-migration. Further, regular fires or human intervention into the vegetational system would invalidate this assumption. To some extent these factors can be controlled for. As long as a full pollen diagram is constructed as well as the pollen ratios, periods of vegetational instability can be identified. Long term successional change will be marked by changes in the nature of the vegetation community represented by the pollen diagram. Frequently, anthropogenic forest disturbance is associated with peaks in 'cultural' pollen types. For example in Europe *Plantago* and *Urtica* pollens are typically associated with periods of human occupation (Behre, 1986). This potential source of error can therefore be identified. Kearney (1983) recommends that surface samples should not be taken from areas with a recent fire history and that fossil pollen studies should be conducted in parallel with counts of charcoal to provide long term fire histories so that decreases in regional *Pinus* pollen are not falsely interpreted as periods of higher treeline (higher *Picea:Pinus* ratio). By careful consideration of parallel lines of evidence, it is possible to obtain a clear picture of the extent to which a particular study site conforms to the assumptions about vegetation stability inherent in the pollen ratio approach.

The influence of varying lake size on pollen ratios in surface sediment was apparent from Maher's (1963) data. The application of surface ratio/elevation data to downcore pollen variation therefore rests on two assumptions about lake size. Firstly that the lakes on which the surface sediment calibration is based on are broadly comparable in size. Obviously this is dependent on the availability of suitable lakes in the study area. The second assumption is that the size of the lake from which fossil pollen is being analysed has not changed dramatically. For deep steep-sided lakes this is a reasonable assumption. However in lakes occupying shallow depressions changes in inputs or outputs from the lake which change lake level, may cause significant changes in lake area. If there is independent evidence of changing lake level (such as past shore line, changing sediment character, etc.) in shallow basins caution should be exercised in interpreting changing pollen ratios. Recent work by Fall (1992) necessitates the addition of a further caveat. Fall demonstrated that moss polsters, pollen traps, bogs, and small lakes without surface inflows record pollen deposition which strongly corresponds to local vegetation. Large lakes however, particularly those with inflows have distorted pollen assembleages, particularly of alpine species. This is due to the large amount of pollen concentrated in such lakes by slopewash. Therefore basins with radically different characteristics should not be mixed in surface sample data sets used to calibrate fossil pollen. In Maher's original work, he noted that the relation between *Picea/Pinus* ratio and elevation differed between large lakes and small lakes or moss polsters although the two lines had similar slopes. Substantial variation in pollen ratios may also be noticed if, for example, lakes which had no surface inflow during the early Holocene climatic optimum subsequently developed inflows in the wetter Neoglacial interval.

Other assumptions of the pollen ratio technique have been identified in addition to those originally outlined by Maher. For example Beaudoin (1986) noted that the application of the technique requires that the slopes over which treeline is fluctuating are relatively uniform in angle. What pollen ratios actually measure is changing deposition of extra-local pollen of treeline origin as treeline fluctuates. This is determined by the changing distance from treeline to the depositional site. However this changing distance is interpreted as representing a changing elevation. Therefore, unless there is a constant linear relation between distance and elevation, i.e. a constant slope, fluctuations in pollen ratios will not truly represent change in treeline altitude.

A further assumption, which Maher does not identify as such but which is implicit in his analysis, is that pollen transport conforms to some regular distance decay function. Maher (1963) does however cite Colwell (1951) to support the suggestion that this function is log linear. Yet the suggestions by Markgraf (1980) and Solomon and Silkworth (1986), that atmospheric inhomogeneities in mountainous terrain affect pollen transport, cause the assumption to be questioned. If the upslope distribution of pollen is described by a 'step function' it will not be possible to interpret changing fossil pollen ratios as a continuous record of changing treeline elevation. This observation further emphasises the need for information from local surface sample transects in order to properly interpret mountain pollen diagrams.

The pollen ratio technique also involves a set of assumptions about the nature of climatic control on pollen production at treeline, considerations of which are critical to correct interpretation. Time lags are likely to exist between shifts in treeline and changes in pollen production. For example, if treeline advances by the establishment of seedlings in the alpine zone, the pollen response will be delayed until the seedlings mature and produce seed. However, if there is a substantial krumholz zone, this effect will be moderated since mature krumholz which have been stressed to the point where pollen production is minimal may start to produce pollen.

The pollen ratio technique assumes that changes in pollen ratios are produced by shifts in vegetation zones. However changes in ratios could be simply due to changing pollen production associated with climatic change. Markgraf (1980) noted decreasing pollen production for individual taxa with increasing elevation, independent of vegetation density. Hall (1990) demonstrated that marked yearly variations in pollen deposition occur. There is no direct evidence that such variation in time and space is associated with climatic control. However Frenguelli *et al.* (1991) showed that the timing of release of *Alnus* and *Populus* pollen in Europe is related to temperature. Also Tranquilini (1979) showed that at high altitudes, dry matter production by vegetation is severely limited by climatic factors. It seems likely that under such conditions pollen production would be curtailed. If it is the case that pollen production is reduced under adverse climatic conditions, then it is possible that observed variations in pollen ratios are a function of change in production rather than distance from source. This phenomena is only likely to be significant over short temporal scales since if treeline is limited by climate, then climate at treeline, and therefore pollen production will be constant at longer time scales. Where fossil pollen is extracted from sediments with a slow sedimentation rate, and hence low temporal resolution, this source of variation may therefore be discounted. However it may be that if high resolution analyses are attempted, high frequency ratio fluctuations will simply reflect changing production. These fluctuations will be a reflection of short term climate variation, but not related to changes in treeline altitude.

2.3.5 Summary Assessment of the Pollen Ratio Approach

The discussion above demonstrates that problems of interpretation of pollen ratio data may arise if the investigator is not constantly aware of the limitations of the technique. It is however undoubtedly a useful approach, producing results which are compatible with other regional environmental reconstructions (e.g. Kearney and Luckman, 1983; Beaudoin, 1986). Pollen ratio analysis is apparently a robust technique. In any one situation practical constraints make it unlikely that all potential sources of error can be eliminated. Despite this the technique produces useful results. Adam and West (1983) applied the technique to a core representing 130 ka of sedimentation in Clear Lake northern California. They showed that treeline shifts parallel the oxygen isotope variation in core V28-238, despite the fact that the assumption of stable vegetational assembleages is not tenable for a period of 130 ka.

With respect to long time spans, and major climatic shifts, the technique is apparently robust. However, close adherence to the assumptions of the technique is probably much more important for studies focussing on shorter timescales and smaller climatic fluctuations.

Having established that under suitable conditions the pollen record may yield a quantitative estimate of former treeline altitudes it is necessary to consider the nature of such evidence. As discussed above the climatic interpretation of treeline elevation data is dependent on the nature of the controls on treeline. In addition. the relation between climate and treeline shift will be subject to a temporal lag due to the time required for tree growth during periods of climatic amelioration. In a study which is attempting to identify small and relatively brief climatic oscillations, it is therefore necessary to carefully consider the potential resolution of palynologically derived treeline data.

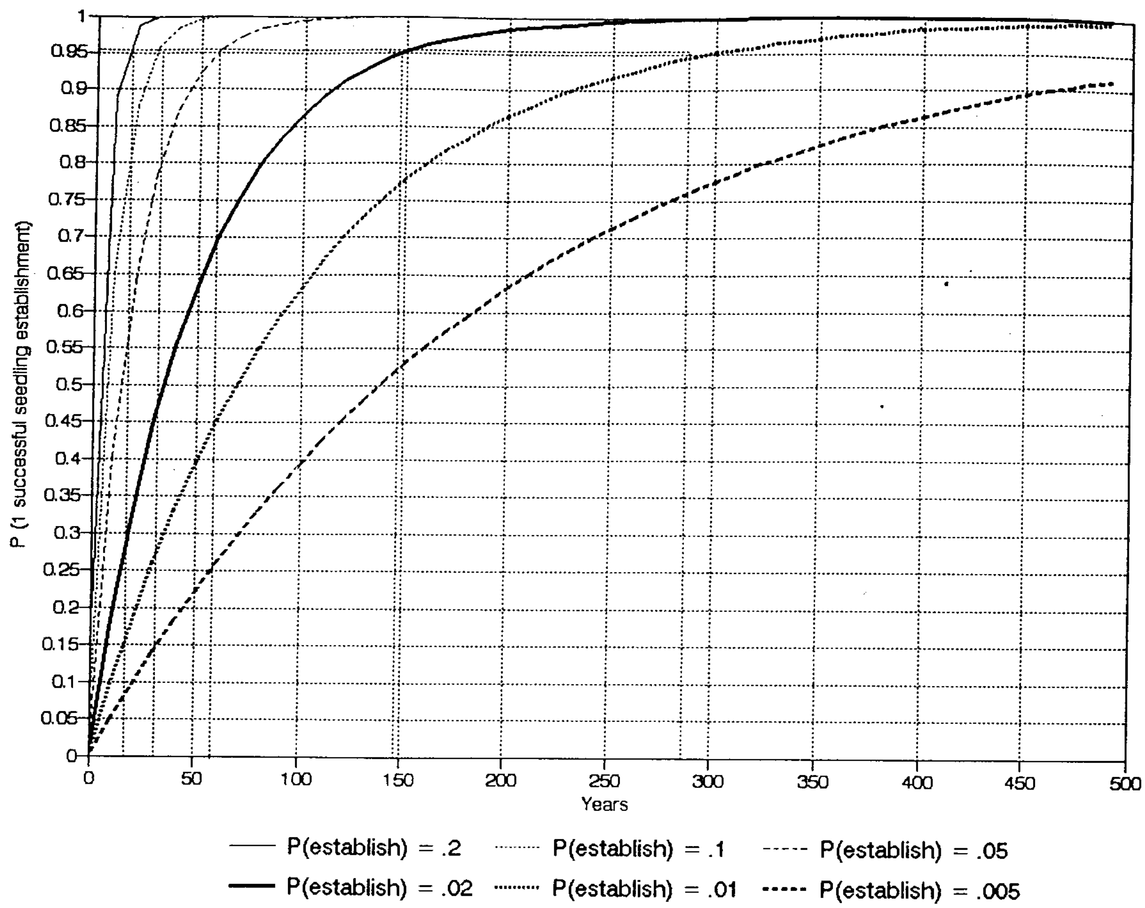
2.3.6 Potential Resolution of Pollen Ratio Data

If quantitative reconstructions of temperature are to be made from pollen ratio data then it is critical to establish the potential resolution of the technique. Otherwise there is the risk that noise will be interpreted as meaningful temperature fluctuations. It is therefore necessary to identify lags in the response of the pollen ratio record to temperature change.

It has already been noted that year to year changes in pollen ratio records may simply reflect variable pollen production. In addition as treeline advances there may be a delay in the increase in pollen production whilst seedlings mature. If however previously stressed krumholz begin to pollinate this lag may be minimal. If reproduction at treeline is halted by climatic deterioration, there may still be pollen produced above the equilibrium treeline by standing trees. A lag equivalent to the lifespan of the tree species in question is therefore introduced. It is possible that this effect too is minimised since there may be a dramatic decline in pollen production by the established trees due to climatic stress.

A potentially more important source of delay between climatic change and pollen response can be identified statistically. As climate warms, there is not a sudden change from conditions where trees may reproduce to conditions where they cannot. Instead there is simply a declining probability that a run of sufficiently good weather to allow seedling establishment will occur. In order to maintain treeline at a particular elevation it is not necessary for reproduction to occur every year. In fact two reproductive years in the lifespan of the tree species will maintain forest cover. Figure 2.1 shows the number of trials required to produce a 95% chance of one success for varying probabilities. This demonstrates that at the altitude where there is only a 1 in 50 chance of seedling establishment, one reproductive year could be expected in 150

Figure 2.1 Period Required for One Successful Seedling Establishment Year, for Various Probabilities of Establishment.



Probability of establishment is a composite measure. Establishment requires that the trees set seed in a given year, that the seed is transported to suitable growth site, and that climatic conditions are sufficient for establishment and survival in that year and subsequent years.

years (95% confidence). If the chance of seedling establishment is reduced to 1 in 100 years then one success might be expected in close to 300 years. This approaches the lifespan of the trees. 0.02 probability of seedling establishment in a given year is therefore an approximation of a critical value. During major climatic changes the probability of seedling establishment just above treeline may move from 0 to 1. However at some elevation temperature will decline so that the probability is around 0.02. This level approximates the theoretical maximum elevation of treeline. However as noted there is likely to be a lag of around 150 years (or any period up to this) before treeline reaches this altitude.

It would appear that the achievable resolution of the technique is something between 0 and 150 years. However on any particular occasion it is not possible to state the exact resolution. The achievable resolution may differ for periods of treeline advance, and for periods of retreat. The resolution may also be further limited by the sedimentation rate and degree of bioturbation in the sediment from which the fossil pollen is extracted. As a rule of thumb therefore it is not reasonable to interpret temperature fluctuations of less than 150 years based on pollen ratio evidence.

2.4 Summary

It is suggested that fluctuations of Neoglacial climate indicated in the glacial record are absent from the pollen records to date because of the lowland nature of those records. A high altitude record might identify minor fluctuations of Neoglacial climate. The altitude of the treeline ecotone is subject to multiple controls. However careful site selection could allow interpretation of treeline fluctuations in terms of local summer temperatures. The sharp gradient of pollen production between the alpine and sub alpine zones makes interpretation of high altitude pollen diagrams difficult. Up valley transport of lowland pollen is a particular problem. Examination of pollen ratios may provide an interpretable and potentially quantifiable record of treeline altitude. With careful adherence to the assumptions of this technique it should provide sufficient resolution to identify the climatic fluctuations of interest in the present study.

Chapter III Study Site

3.1 Study Site Selection

Site selection for palynological studies has been regarded as something of a "black art". The emphasis has tended to be on the identification of long, high resolution records. In general, the nature of the record likely to be preserved at different sites due to their geographical location has received less attention. In part this is due to the fact that many studies have been aimed at the reconstruction of regional vegetation patterns in Holocene time. The balance between the regional and local components of pollen rain is controlled largely by basin area (Jacobsen and Bradshaw, 1981). Therefore, for regional studies, as long as the basin is of sufficient size, geographic location is of secondary importance. The limited spatial distribution of sediment sequences which preserve a palynological record is also a constraint on geographic considerations in site selection. The aim of this study is to identify local changes in the vegetation pattern at treeline. This necessitates sampling from a relatively small basin. Jacobsen and Bradshaw (1981) show theoretically that around 75% of pollen recruitment is of extra-local or local origin for a basin of 200 m diameter. The relation between basin size and pollen recruitment assumes a lake with no inflow. For lakes with an inflow Jacobsen and Bradshaw suggest that the representation of streamside communities in the pollen spectra will increase. For alpine sites close to the head of stream system this amounts to increased local pollen influx. In small alpine basins therefore studies of local pollen may need not necessarily be limited to lakes without an inflow. Emphasis on local change however necessitates a careful consideration of the geographic location of the site.

Hence, the primary study site requirements for this study of treeline fluctuations were a small (< 200 m diameter) lake or bog at treeline providing a continuous sedimentary record of the Neoglacial period. The site should therefore be outside any Neoglacial trimlines. In addition, since sedimentation rates in closed alpine lakes may be very low (Owens, 1990), lake sites should have an apparent inflow and sediment source. Such a site should provide a resolvable palynological record of treeline fluctuations. However the primary intent of this study is not to examine Neoglacial treeline fluctuations but to interpret such fluctuations in terms of changing climate. Further constraints on site selection are therefore necessary. As noted previously, temperature may provide a theoretical maximum treeline altitude, but this can be modified locally by a variety of factors. At any one site it is not possible to state categorically the environmental determinant of treeline altitude without detailed study of local modern ecology. In an attempt to isolate a site where temperature is the primary determinant of treeline altitude, the following selection criteria were established. These rules are based on observations of potential sites, analysis of air photographs of the area, and available data on regional climate.

1) Treeline should be at a relatively constant altitude in the local area.

Mean thermal regimes display little small scale spatial variability relative to other potential controls on treeline (soils, moisture, snowpack, insect infestation, wind, fire). Therefore, if temperature is the controlling variable, treeline altitude should not vary much within a local area.

2) There should be no apparent edaphic limitations.

For example treeline should not be depressed because scree or glacial trimlines prevent establishment at higher altitudes.

3) No "excessive" snowpack.

Late lying snow may inhibit seedling establishment and shorten the growing season.

4) A site where treeline is not unusually low in the regional context.

As noted above, temperature controlled treelines are at the theoretical maximum altitude. Temperature controlled treeline sites are therefore likely to have the highest local treeline altitudes.

Once criteria for site selection were established, maps and air photographs of the southern Coast Mountains were examined for possible sites. In much of this region, treeline is depressed by extremely thick winter snowpack, which lies late into the summer (Arno, 1990). Therefore, site selection was narrowed to the eastern flank of the mountains close to the transition between the Coastal and Interior Douglas Fir Biogeoclimatic zones. Initially, two possible sites were identified. These were two lakes at the head of Blowdown Creek (Grid reference NTS 92 J8 600790), Blowdown Lake and Upper Kidney Lake; and Upper Joffre Lake (Grid reference NTS 92 J8 373772).

These lakes were reconnoitered in early June 1992. Surface samples of lake sediment were taken with an Ekman grab sampler deployed from a small zodiac dinghy. Preliminary bathymetric data were recorded with a small echo sounder. At Upper Joffre Lake this revealed the presence of a submerged terminal moraine which divides the basin into two. This marks the maximum Little Ice Age extent of Matier Glacier which lies above the lake. Approximately five centimetres of sediment was recovered from the outer basin which was a grey inorganic silt/clay. The lake is steep sided with a depth in the centre of around 30 metres. With the available equipment, coring in 30 metres of water would have been impractical. Upper Joffre Lake was therefore rejected as a study site. It is however a site worthy of further study. If cores could be obtained from both sub basins, perhaps using a gravity corer, a detailed study of the effects of the little Ice Age on lake sedimentation might be possible.

At Blowdown Creek, the Ekman sampler recovered 20 cm of soft organic gyttja from both lakes, suggesting a useful record. The site met the criteria established above, and was therefore selected as the study location.

3.2 Blowdown Lake

Blowdown Lake is a small (200 m diameter), oligotrophic lake at treeline (elevation 2025 m)(Figure 3.1, 3.2). It is located to the northeast of Lillooet Lake, and southeast of Duffey Lake, in the southern Coast Mountains of British Columbia (Figure 3.3). The lake is located in a large corrie at the upper end of Blowdown Creek. There are two streams flowing into the lake. The main input at the east end of the lake is a steep, straight rocky channel running down from a small pond on the ridge above. A second smaller stream flows in from the north. The south side of the lake is bordered by a steep talus flanked cliff with at least one well developed debris chute. The cliff exhibits considerable development of tension cracks. The slopes to the north of the lake are long straight and covered with herbaceous vegetation and a few krumholz patches. The vegetation has been stripped in many places and the slopes appear to be rapidly eroding (Figure 3.4). Rocky debris, and broken and tilted trees at the edge of the lake (Figures 3.5, 3.6) suggest that avalanches reach the lake. The bathymetry of the lake was established by rowing transects across the lake whilst operating an echo sounder. Distances were determined using a hip chain. The bathymetry is represented in Figure 3.7.

Figure 3.1 Blowdown Lake

Figure 3.2 View of Blowdown Lake from Above

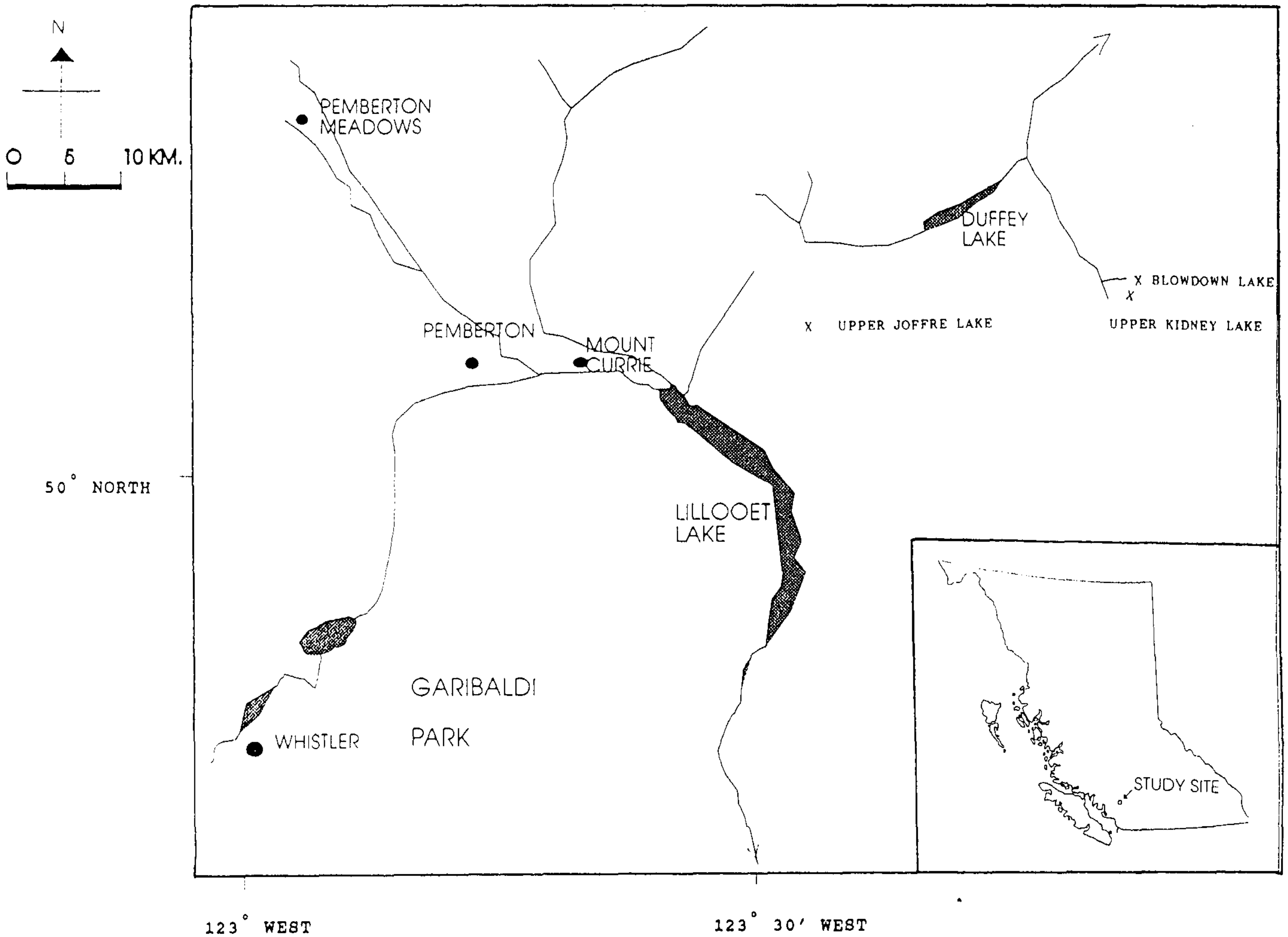


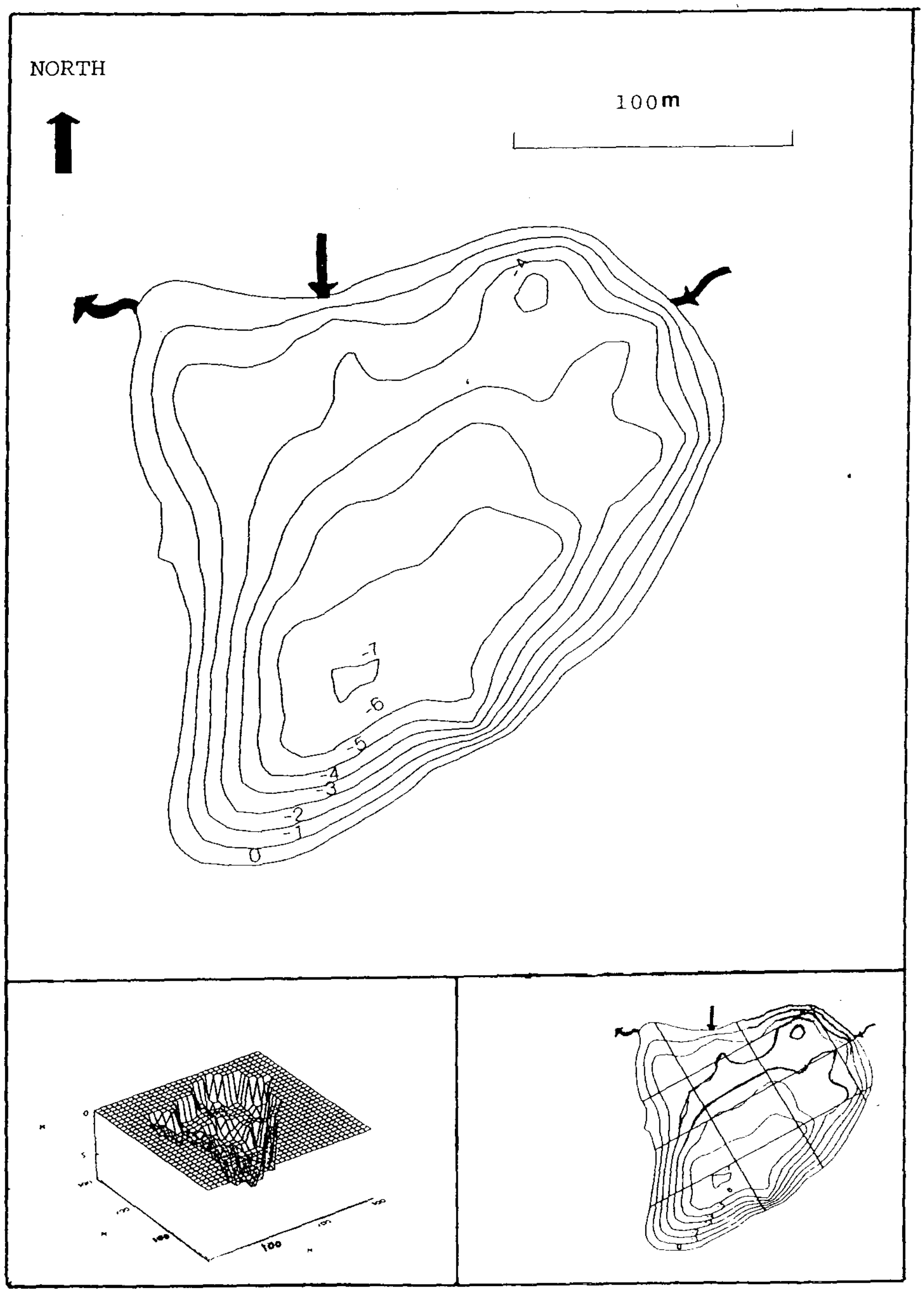
Figure 3.3 Sketch map of the Location of Lakes Mentioned in the Text

Figure 3.4 The Slopes to the North of the Lake

Figure 3.5 Avalanche Debris on the North Side of the Lake

Figure 3.6 Tilted Trees next to the Lake

Figure 3.7 Bathymetry of Blowdown Lake



Contour interval in metres. Left inset, three dimensional representation of the bathymetry. Right inset, Sonar transects on which the contouring of bathymetry is based. Sonar readings were taken every 10 metres along the transects.

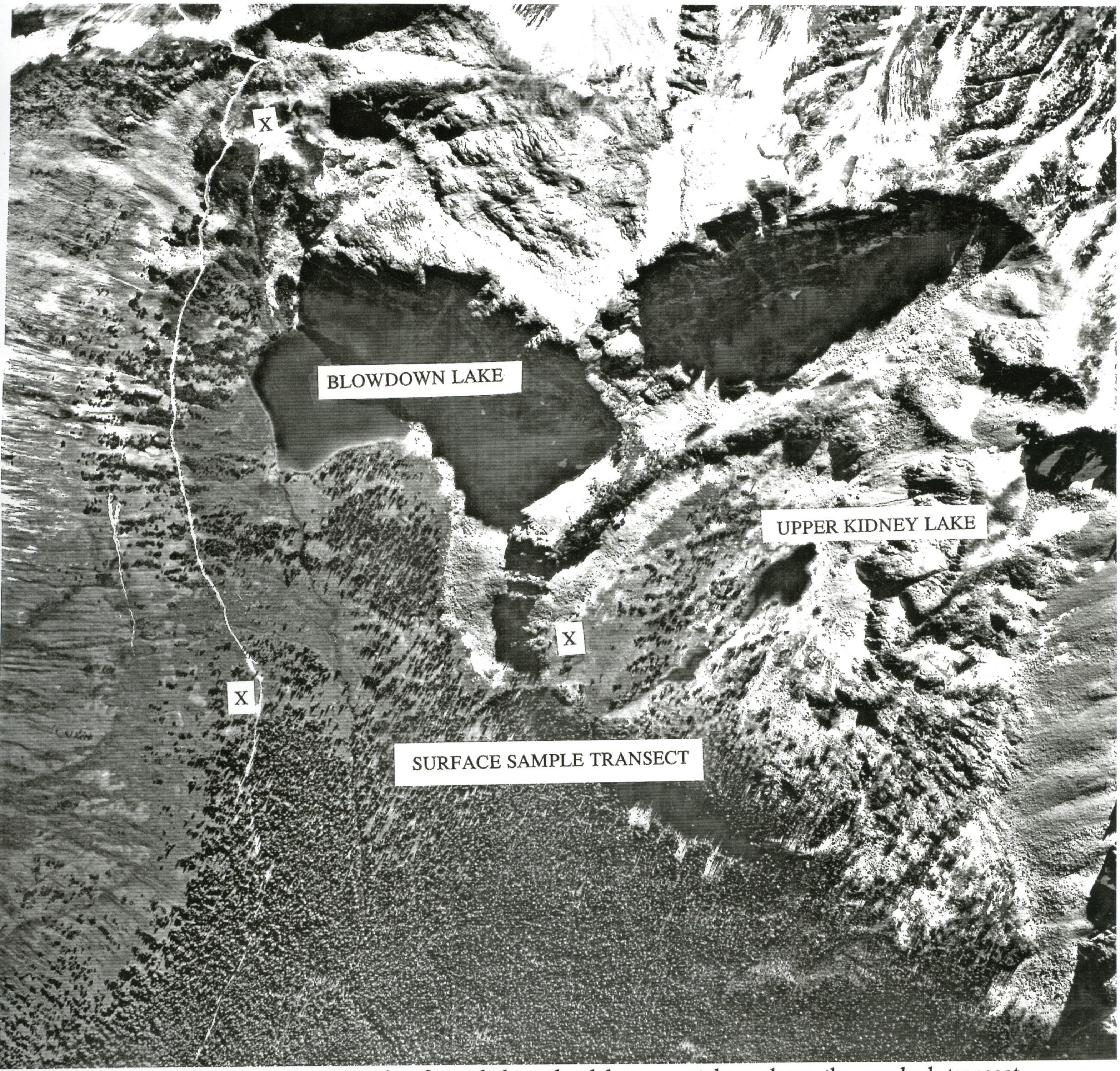
3.3 Upper Kidney Lake

Upper Kidney Lake is separate from Blowdown Lake by a rocky ridge to the south of Blowdown Lake (Figure 3.8). It lies at an elevation of around 2050 m in an overdeepened corrie. On both sides of the corrie, rock walls have developed considerable talus slopes. A lobe of bouldery material, perhaps a small rock glacier flows into the corrie over the headwall. There is a single inflow from the east and an outflow at the south end. The bathymetry of the lake was determined in the same manner as at Blowdown lake, the results are presented in Figure 3.9.

3.4 Climate

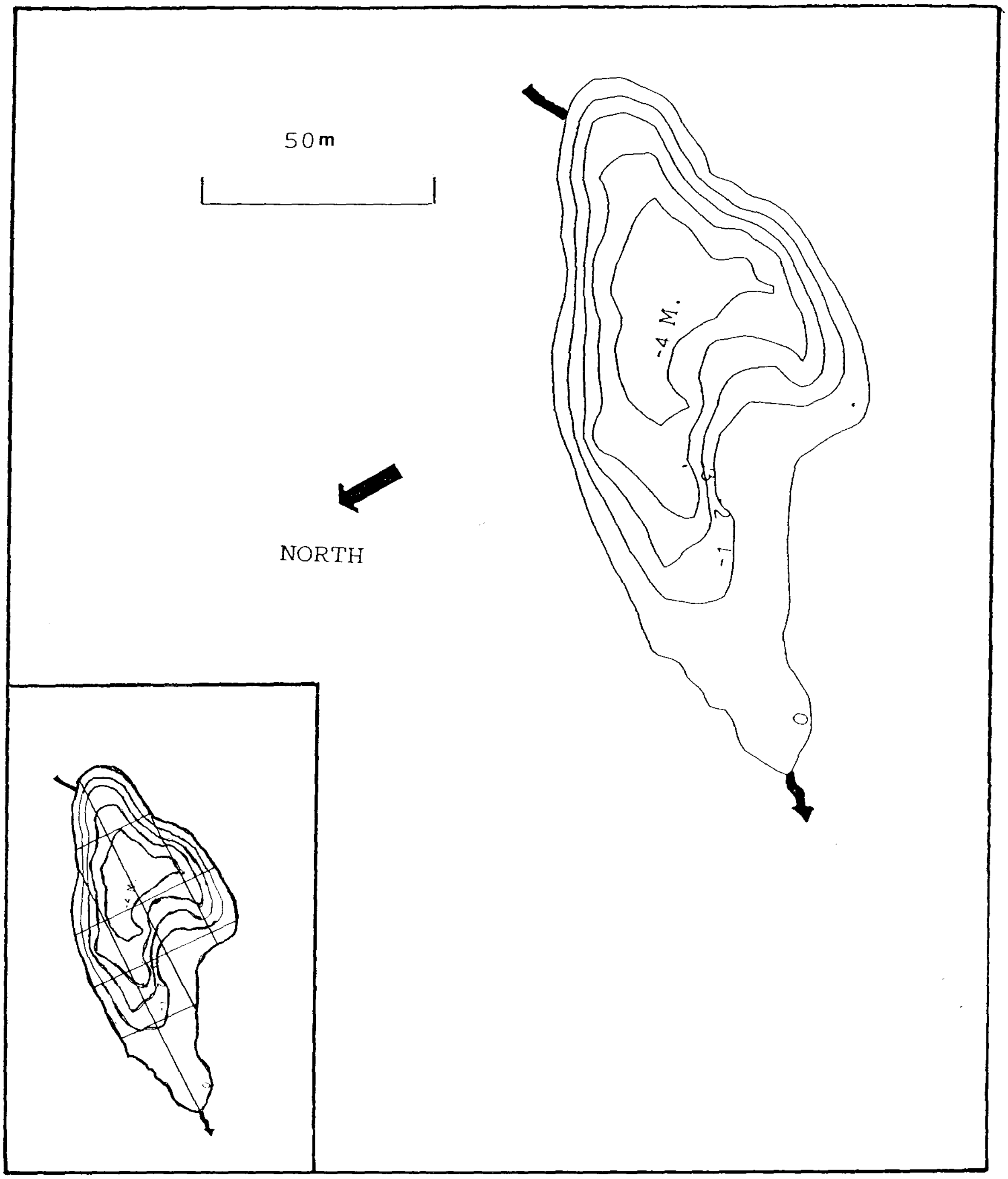
No weather data are available for the site. In 1992 the site was clear of snow by June 1 although this was a low snowpack year. In 1993 the state of vegetational development on June 25th suggested that the site was very recently clear of snow. To the north of the lake behind a ridge are a series of small rock glaciers. At least one is appears active since several concentric lobes of material are present flowing away from the cliff. This would suggest that the mean annual temperature of the site is close to 0°C. The closest weather station to the lakes is in Pemberton, 45 km to the west. The Pemberton data, taken at an elevation 218 m, are not representative of the climate at 2025 m. Perhaps a better approximation to the local climate is data from Goat Meadows (Gallie, 1983; Gallie and Slaymaker, 1984, 1985; Owens, 1990) Goat meadows are located near 1700 m. in the Coast Mountains above Pemberton Meadows, 50 km west and 20 km north of the site. Data from this site are recorded in Table 3.1. The range of temperature at this site may be similar to

Figure 3.8 Air Photograph of the Headwaters of Blowdown Creek



Nine surface samples from below the lake were taken along the marked transect. Samples from elevations above the lake were taken at the sites marked with an x. (See 4.1.2)

Figure 3.9 Bathymetry of Upper Kidney lake



Inset map marking sonar transects upon which the bathymetric contours are based. Sonar readings were taken every 10 metres.

Table 3.1 Climate data from the Goat Meadows Watershed

Annual precipitation	> 1800 mm
	70 - 80% as snow
Mean annual air temperature	0 -1°C (1979 figure)
May 1 Snowpack	1500 mm water equivalent

Data taken from Owens, 1990; Gallie, 1983. Data are estimates based on comparison of 1979-1980 data with longer records from nearby snow courses and weather stations.

Blowdown Creek, however the precipitation is probably somewhat higher because the Goat Meadows site further west close to the Lillooet icefield. In contrast, there is no evidence of Holocene glaciation in the vicinity of the Blowdown sites, although permanent snow patches may have developed during the Neoglacial phases.

3.5 Geology and Surficial Materials

Geology for the Pemberton (92 J) map area has been mapped by Woodsworth (1977). Blowdown Lake lies on the border between areas of quartz diorite and granodiorite. The granodiorite cliff above the lake appears to mark the boundary. The Kidney lakes lie in the granodiorite zone. The site is partially covered by the typical sandy till of the Coast Mountains, containing abundant boulders. Soil development is minimal, although a humic gleysol has developed in the wet meadows below the lake. In the wettest areas thick organic soils are found.

3.6 Vegetation

The lakes lie on the transition between the Coastal Mountain Hemlock biogeoclimatic zone, and the interior Englemann Spruce - Subalpine Fir zone. The local vegetation at the site is typical of the treeline ecotone in this region. There is a transition from subalpine forest to a parkland of scattered tree islands. The forest is dominated by *Abies lasiocarpa* and *Picea engelmanii* along with the occasional *Pinus monticola*. The same species compose the parkland tree islands although *Abies lasiocarpa* is dominant. The parkland understory is highly differentiated, the principlal control being moisture variability due to microtopography. Dry sites are dominated by *Cassiope mertensiana*, and *Phyllodoce empetriformis*. These two species along with *Pinus monticola* are dominant around Upper Kidney Lake which is a much drier site than Blowdown Lake.

In wetter areas the shrub understory is *Alnus sinuata*, and two species of *Salix*. The area below Blowdown Lake is a wet sedge meadow supporting various *Cyperaceae* and *Gramineae*, *Salix spp.*, *Menziesia ferruginea*, and abundant alpine herbs. A species list of herbs and shrubs, compiled whilst walking from the meadow past the lake to the ridge above, is presented in Figure 3.10. Above the lake, vegetation is more stunted with krumholz forms of the three tree species at the site, *Gramineae*, and various alpine flowers. At the highest elevations *Juniperus communis* is an occasional shrub.

Seedlings of *Picea* and *Abies* were observed in clearings in the forest below the lake. Regeneration does not appear to be taking place around the lake, although fertile *Abies* were observed. Above the lake, krumholz reproduction appears to be primarily vegetative. Creeping layered forms, particularly of *Abies*, were observed.

Figure 3.10. Blowdown Basin Species List. Nomenclature follows Hitchcock and Cronquist (1973)

Alnus sinuata
Salix spp.
Menziesia ferruginea
Vaccinium caespitosum
Phyllodoce empetriformis
Cassiope mertensiana
Ranunculaceae
Equisetum arvense
Valeriana sitchensis
Veratrum viride
Heracleum lanatum
Castilleja spp.
Silene acaulis
Habenaria dilatata
Erigeron spp.
Veronica alpina
Lupinus arctica
Antennaria spp.
Pedicularis bracteata
Anenome occidentalis
Kalmia polifolia
Claytonia lanceolata
Thalictrum occidentale
Phlox douglasii
Viola spp.

Chapter IV Fieldwork and Laboratory Methods; and Presentation of Results

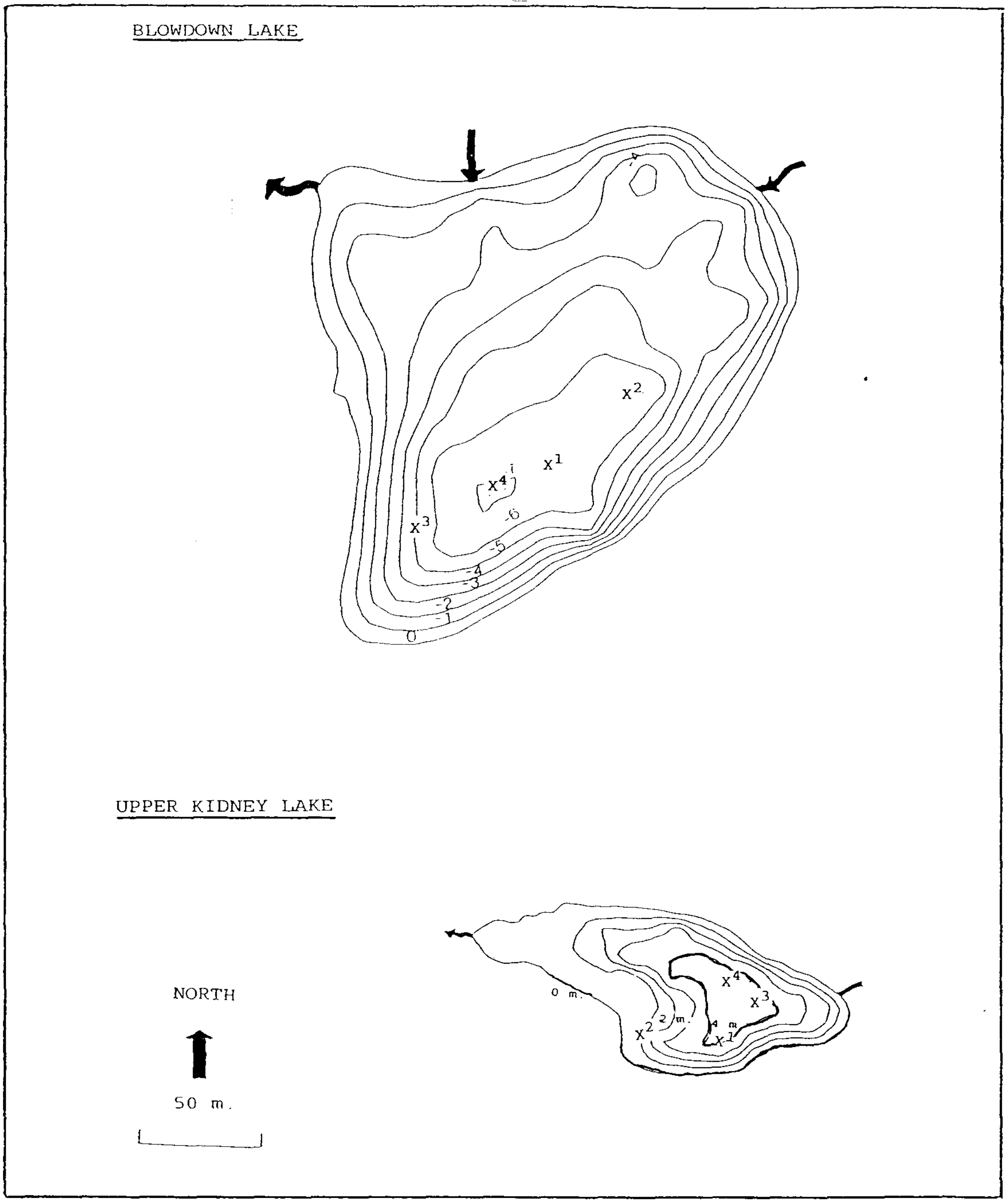
4.1 Field Methods

4.1.1 Lake Sediment Sampling

Lake sediment cores were collected from Blowdown and Upper Kidney Lakes during late June 1992. Coring was carried out from a raft constructed for the purpose (Figure 4.1), using a modified Livingstone corer (Livingstone 1955). The raft was anchored with rocks at each corner over the coring site. Coring was conducted inside a casing of black ABS pipe. The coring operation was not straightforward because the sediments in both lakes are relatively stiff. In particular, coarse sand layers were prevalent in Blowdown Lake. Even with anchors at each corner, because of the substantial water depth, the raft was not particularly stable. The coring rods tended to flex and then straighten driving the raft away from the vertical position over the core site. This reduced the force which could be brought to bear on the corer. The result of these difficulties was that systematic coring of the lakes was not possible. Cores could only be obtained where thick sand layers or rocks did not impede penetration. Blowdown lake presented particular difficulties since the bed is littered with large boulders. Usable cores were extracted from several sites in both lakes (Figure 4.2). These were extruded on the raft. Due to the sandy nature of the sediments, it was not possible to extrude the cores by hand. The corer does not perform well in sandy sediment (Schmolk, 1986). A ratchet device with a mechanical advantage was therefore used to extrude the core. This involved considerable compression of the core, particularly of the wet surface sediments. However the stratigraphy of the cores was preserved. The extruded cores were wrapped twice in clingfilm and then in Thick plastic. the cores were stored

Figure 4.1 Portable Raft Constructed for Coring the Lakes

Figure 4.2 Location of Coring Sites at Blowdown, and Upper Kidney Lakes.



in lengths of plastic piping.

4.1.2 Surface sediment samples

In addition to the lake sediments, surface sediment was also collected in order to provide a record of modern pollen deposition. Nine samples were taken along an altitudinal transect starting in the meadows below Blowdown Lake. Altitude for each sample was recorded using an altimeter. These samples were of wet humus that has accumulated in damp depressions alongside a network of small streams. Sites which did not appear to dry out over the summer were selected. Such sites were easily available in the wet meadow below the lake. Wet sites are rare above Blowdown Lake. Therefore samples were taken from three small ponds on the slopes above the lake. The locations of surface sample sites are marked on Figure 3.8.

4.2 Laboratory Methods and Results

In the laboratory, cores were unwrapped and split lengthways with a sharp knife. The longest cores from each lake were selected for study. The visible stratigraphy was described according to the Troels-Smith classification scheme (Troels-Smith, 1955). Colour was described using Munsell charts. Once the core had been described half was rewrapped and stored. The other half was divided into one centimetre sections. Each section was stored in a sealed plastic petri dish.

4.2.1 Biological analyses

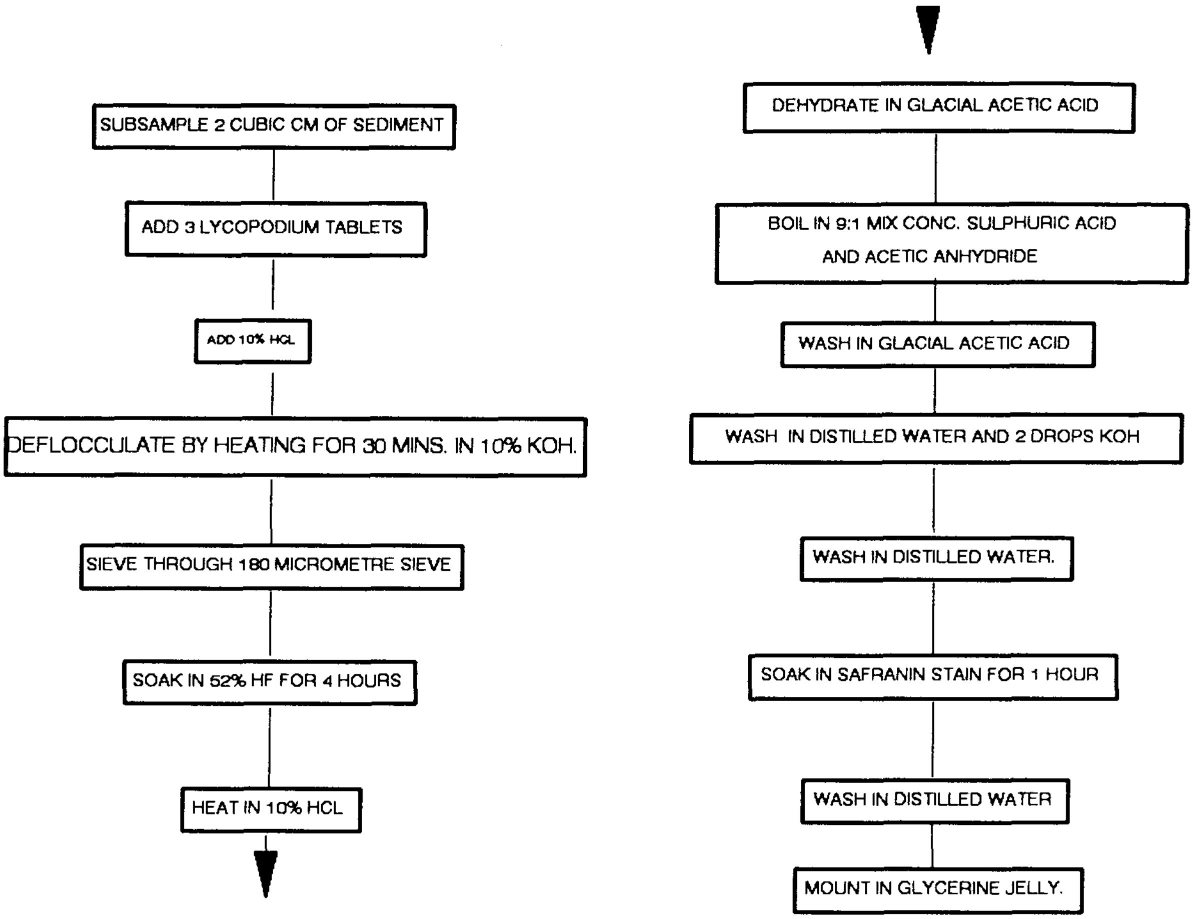
4.2.1.1 Pollen analysis

All the surface samples were analysed for pollen using standard methods (see Figure 4.3). Samples from cores were taken at 5 cm intervals. Two cm^3 of sediment was sampled by removing a plug of material with a graduated syringe. Further samples were taken from stratigraphic horizons of interest once the initial pollen stratigraphy had been established. The pollen preparation procedure followed was essentially that of Faegri and Iversen (1989). In order to allow absolute counting, three commercial lycopodium tablets containing a known number of spores were added to each 2 cm^3 sample. The pollen preparation procedure concentrates the pollen in the sample by chemically dissolving other components of the lake sediment. It is summarised in Figure 4.3.

The pollen slides were counted by scanning on a Zeiss light microscope at 400x magnification, with critical identifications being made under oil immersion at 1000x. Pollen identification was by reference to atlases; primarily Moore and Webb, (1978), Dichotomous keys; Moore and Webb (1978), and an unpublished key provided by Dr Glenn Rouse (U.B.C. dept. of Botany), and to the type collection belonging to Dr Rouse.

As noted in section 3.6 the vegetation of the Upper Kidney Lake catchment is significantly different from that of the Blowdown Lake catchment. This difference is assumed to be due to the thin freely draining soils in the former catchment. Since the surface pollen samples were taken from small humus accumulations, local pollen types will be strongly represented. Given the extent of local variation in vegetation it was regarded as unreasonable to attempt to interpret the Upper Kidney lake pollen record on the basis of surface samples taken from the Blowdown catchment. Because of the drier

Figure 4.3 Summary of Pollen Preparation Procedure



nature of the Upper Kidney Lake catchment suitable sites for taking surface samples were not available. Therefore because of the prospect of calibrating the pollen record, it was decided to concentrate further analyses on the sediments from Blowdown Lake. The results of palynological investigation of Blowdown Lake are presented and analysed in Chapter VI.

4.2.1.2 Needle Fragments

In order to calculate the concentration of needle fragments it is necessary to count needles from a known volume of sediment. Five centimetre sections of half core were taken and volume of sediment ascertained by displacement of water in a measuring cylinder. The volumes of sediment sampled for needles were around 20 cm^3. The samples were then soaked in a 3% solution of KOH overnight to deflocculate the sediments. The resulting slurry was sieved through a 500 μm filter, and washed with distilled water. Needle fragments were picked out with tweezers under the binocular microscope. The fragments were stored in a solution of 70% ETOH.

The fragments were identified and counted under a binocular microscope. Identification was by reference to the key of Dunwiddie (1985) and the reference collection of Dr Rolf Mathewes at Simon Fraser University. During the identification of needle fragments any charring of the needles was noted. Following Wainman and Mathewes (1990) needle fragments were classified as tips, bases, or mid-sections. The number of needles in the sample was calculated as (tips+bases)/2. In this study, this method did not appear to underestimate *Pinus monticola* needles (a problem noted by Wainman and Mathewes).

Once the needles were extracted, concentrations of needle fragments were calculated. In order to obtain sufficient sample for these calculations to be meaningful samples were aggregated. Therefore in the upper half of the core, where needles were scarce, the data presented represent 20 cm of depth of the core per data point In the lower half of the core each sample represents a 10 cm horizon. The results are presented in Figure 4.4.

Wainman and Mathewes, (1990) demonstrated that macrofossil concentrations are higher near shore compared to the centre of a lake. Needles were counted from the main core in the centre of the lake (BD4A), therefore the number of needles of each species in the samples was relatively low. Despite this problem analysis of needles from core BD4A was preferable since it enabled direct comparison with other the analyses.

Figure 4.5 illustrates the results of microscopic examination of needle macrofossils for evidence of charring. The data are presented as the percentage of charred needle fragments compared to all fragments identified in each horizon.

4.2.2 Physical Analyses

4.2.2.1 Stratigraphy

In addition to the initial description of stratigraphy, the intact half of the core was later opened, dried, and then the stratigraphy described again because further stratigraphic detail became apparent during the drying. The core was also photographed to record the stratigraphy.

The intact halves of the cores were taken to the UBC hospital and X-rayed. The automatic phototimer setting of the X-ray machine was used.

The visual stratigraphy of all the cores extracted from Blowdown Lake is presented in Appendix 2. The major stratigraphic features of these cores are the distinct sand layers found in all cores. These are continuous between adjacent cores, although correlation across the whole lake is difficult. The preservation of these distinct layers

Figure 4.4 Concentration of Conifer Needle Macrofossils in Core BD4A

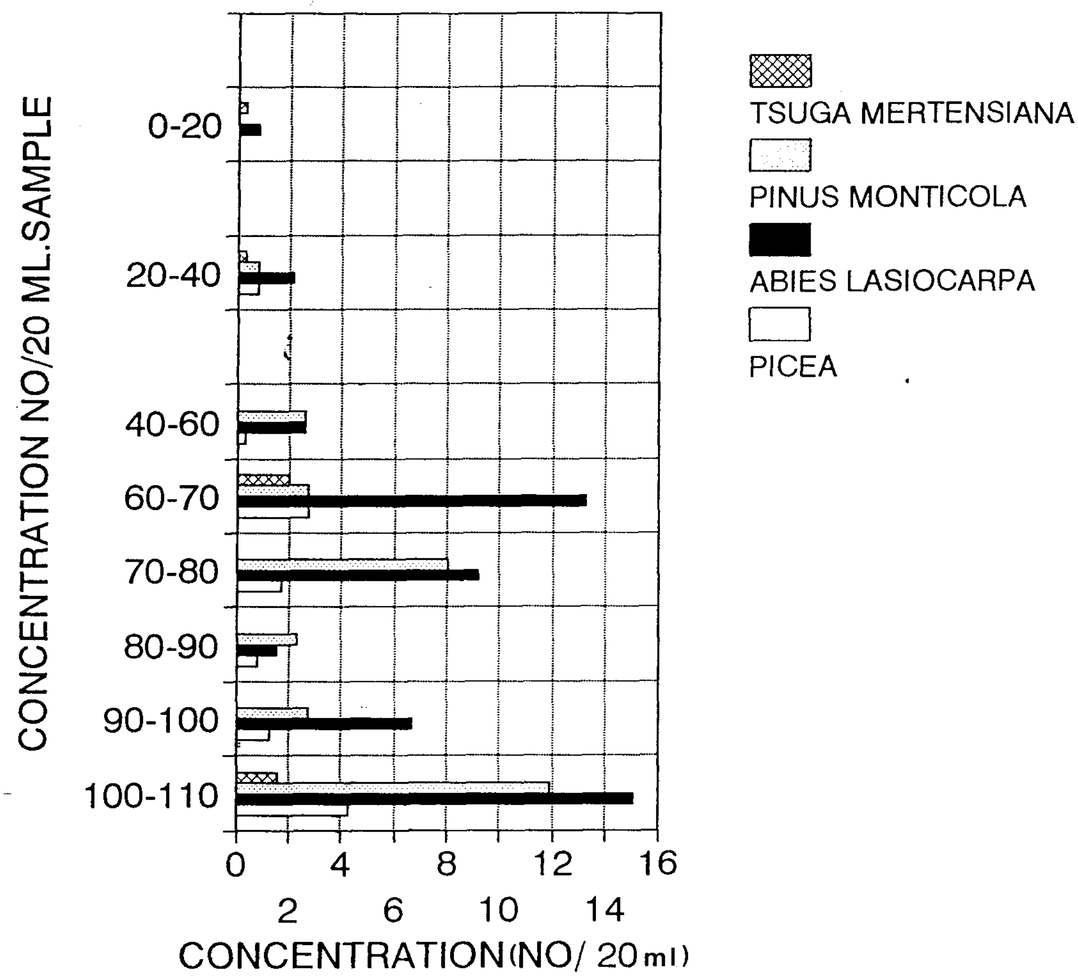
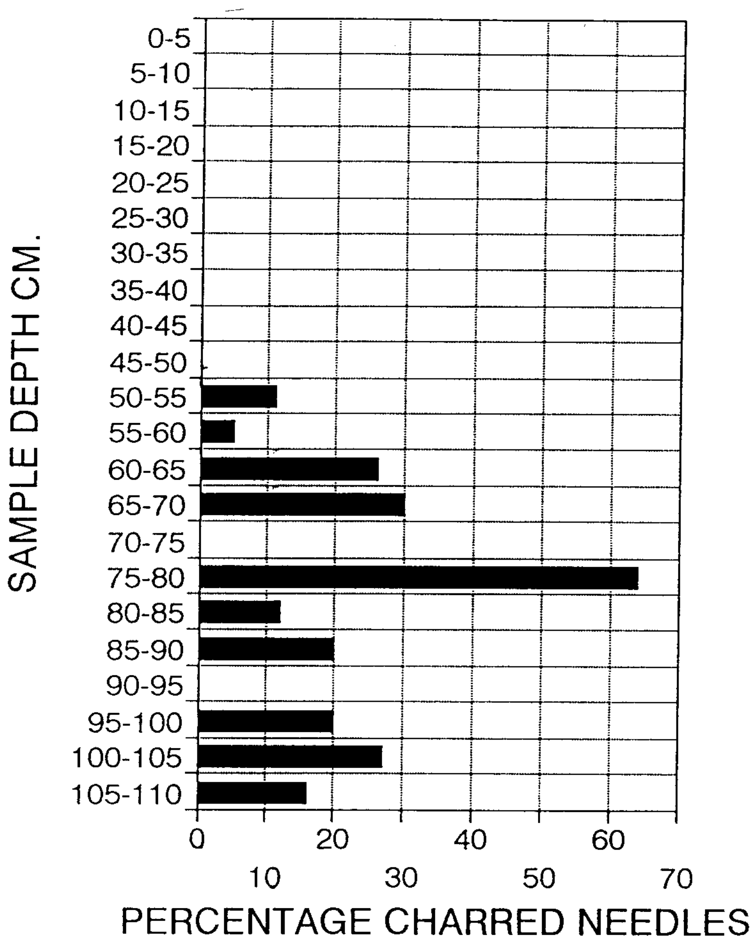


Figure 4.5 Percentage of Conifer Needle Macrofossils with Apparent Charring - Core BD4A



suggests that the stratigraphy of the cores is largely intact. This is important given the presence on the lake floor of large boulders: the distinct sand layers allow some confidence that sedimentation took place around the boulders, and that they did not subsequently fall into the lake disturbing the sediment. The boulders may provide an explanation for the difficulty in correlating the layers across the whole lake. They provide a very inhomogenous surface to the lake floor, such that the spatial distribution of sediment deposition from a single major input might be very variable.

The stratigraphy of core BD4A is illustrated in Figures 4.6 and 4.7. Figure 4.6 is a composite representation of stratigraphic features visible in the core both wet and dried, and on the x-ray plates. The results of Troels-Smith characterisation of the core by visual inspection are presented in Figure 4.8. Three features of the core are of particular note:

1) Sand layers occur throughout the core. Beneath 50 cm depth the layers are thin (1-2 mm), and of fine to medium sand, above 50 cm the sands are thicker, (up to 3 cm) and fine upwards. The thickest layers fine from fine gravel to medium sands, these upper sand layers are also include dropstones, and some slightly disturbed stratigraphy (apparent from the x-rays).

2) The second immediately obvious feature of the core is the colour change from black (2.5Y 3/2) at the base, to greyish brown (10YR 5/2) at the surface. This reflects a change from organic gyttja at the base to strongly minerogenic sediment at the surface. This colour change was apparent in all cores from Blowdown Lake providing further evidence of orderly sedimentation in the lake.

3) The impression of a shift from organic sediment in the lower half of the core to less organic sediment above, is strengthened by the presence of distributed conifer needles in the lower sediment. The needles are most apparent from 60 - 80 cm. and from 90 - 110cm. depth.

Figure 4.6 Core BD4A Stratigraphy

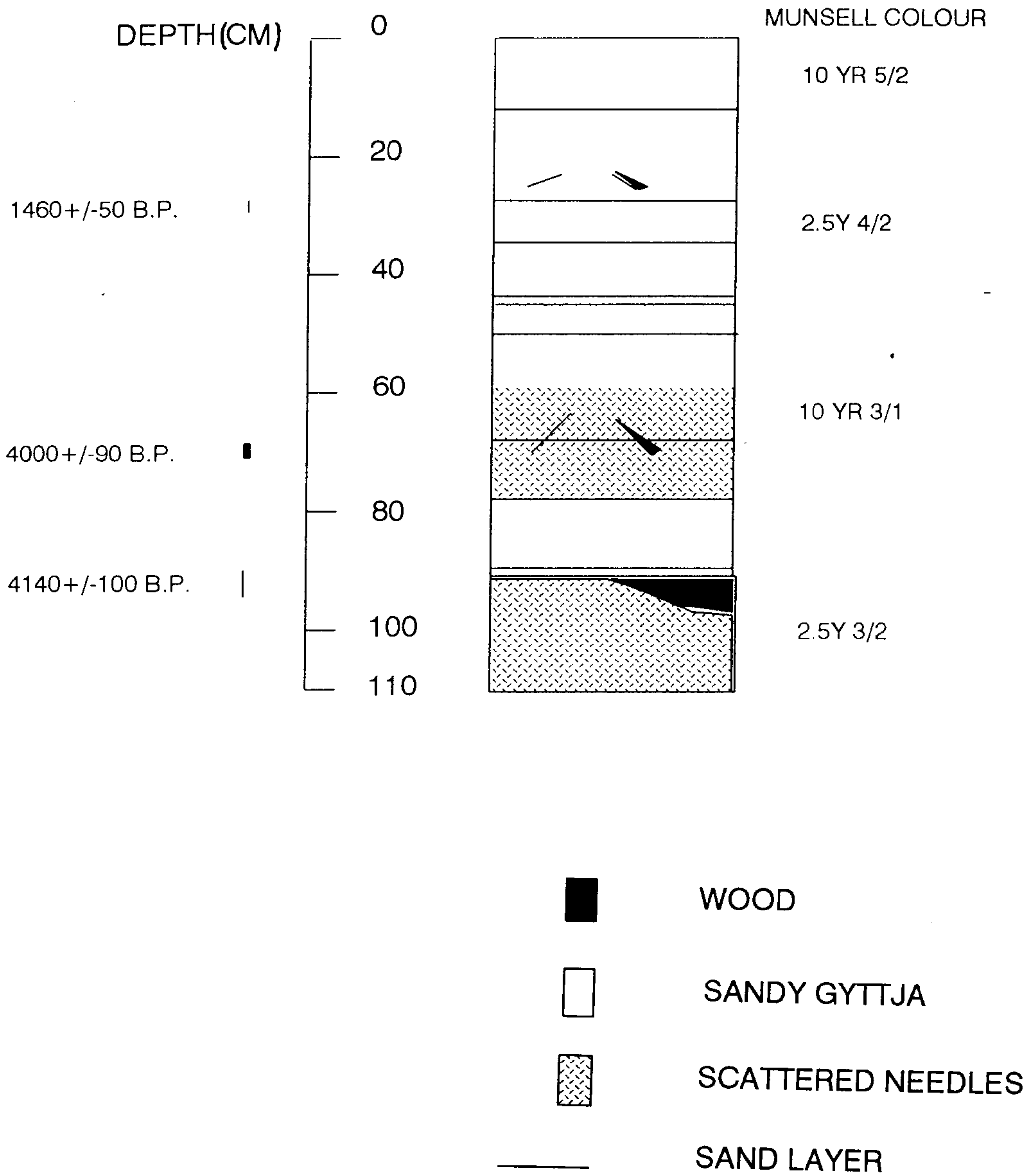
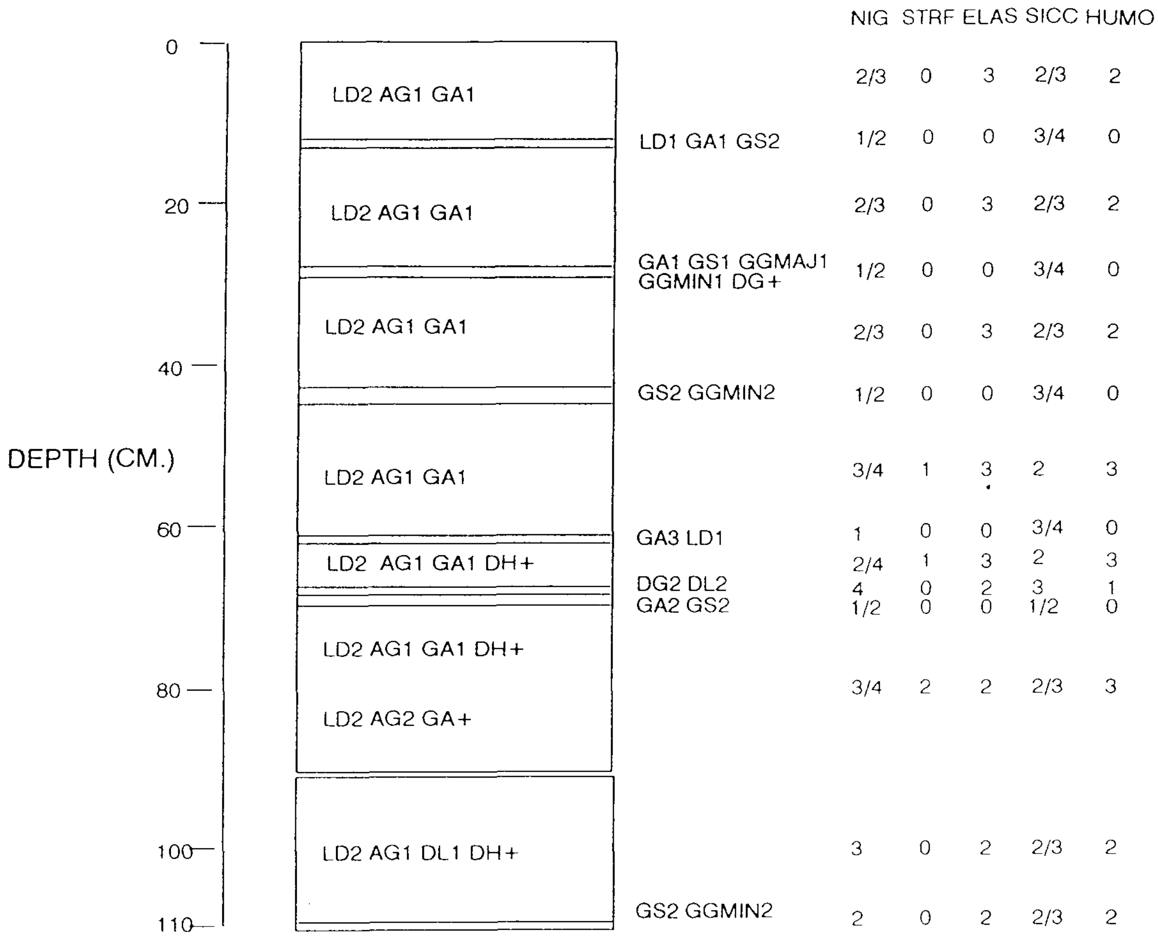


Figure 4.7 BD4A - Photograph of one Half of the Split Core

Figure 4.8 Troels-Smith Classification of Core BD4A



LD - *Limus Detritus* DL - *Detritus Lignosus*
DH - *Detritus Herbosus* AG - *Argilla Granosa*
GA - *Grana Arenosa* GGmin - *Grana Glareosa (Majora)*
GS - *Grana Suburralia* GGmaj - *Grana Glareosa (Minora)*

Nig - *Nigrum* Strf - *Stratificatio* Elas - *Elasticitas*
Sicc - *Siccitas* Humo - *Humificatio*

For details of the Troels-Smith Classification Scheme refer to Troels-Smith (1951) or Birks and Birks (1980).

Three dates are shown in Figure 4.6. The date from 90-96 cm depth is a conventional radiocarbon date on a log, the date at 67 cm is an AMS date on conifer needles, and the date at 28 cm is an AMS date on a small twig. These dates confirm that the core spans at least the majority of the Neoglacial period. A more detailed discussion of core chronology is given in Chapter VII.

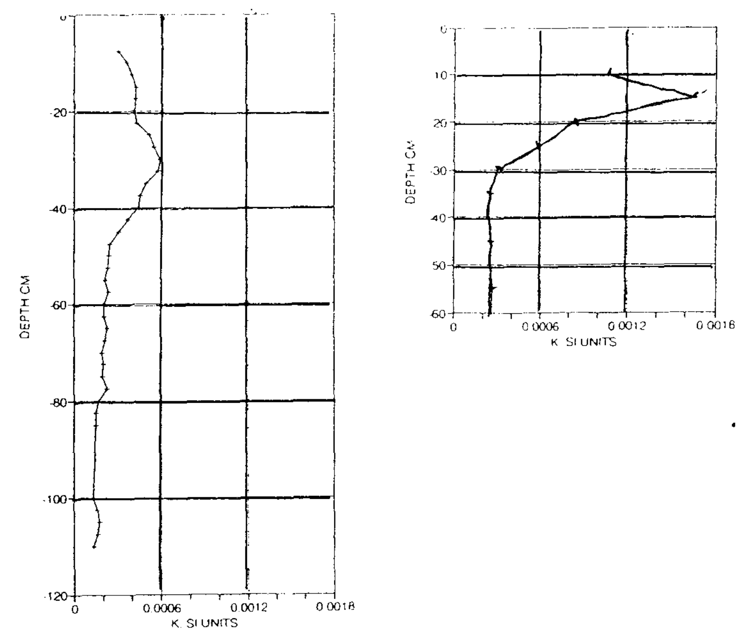
4.2.2.2 Magnetic Stratigraphy

Cores from each lake were taken to the Pacific Geoscience centre at Patricia Bay on Vancouver Island for magnetic susceptibility analysis. By running the whole core through a magnetic coil, readings of magnetic susceptibility (k) were taken at five centimetre intervals. This provided discrete data since the measurement field of the coil was five centimetres. Measurements were also taken at 2.5 cm intervals which effectively provides a moving average of readings of k. This method of whole core scanning is a rapid and non-destructive way to obtain useful stratigraphic data.

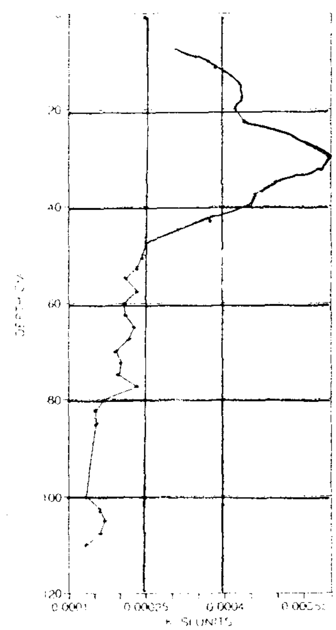
Sediments may be characterised by their magnetic properties in a variety of ways. Magnetic susceptibility is "the ease with which a material can be magnetised" (Thompson and Oldfield, 1986, p.25). Volume susceptibility k is defined as M/H where M is the magnetisation induced by a field H. Volume susceptibility can be shown to be a good approximation of the magnetite concentration in a sample where magnetite is the dominant magnetic mineral (the usual situation) (Thompson and Oldfield, 1986). Thompson *et al.* (1975) demonstrated that susceptibility peaks were consistent in multiple cores in a basin, and thus useful for core correlation. The susceptibility traces for two cores from Blowdown lake are presented in Figure 4.9. It is apparent that both cores exhibit the same broad patterns of magnetic stratigraphy, with a large peak in k towards the top of the core. This evidence of within basin correlation further strengthens the assumption that the stratigraphy of the lake

Figure 4.9 Comparison of Magnetic Susceptibility Values of Two Cores from Blowdown Lake

A)



B)



A) Magnetic susceptibility traces for two cores from Blowdown lake (comparable x axis scales).

B) Magnetic susceptibility trace for the main Blowdown lake core, showing detail.

Overlapping "moving average" data are available for BD4A. These provide more detail than the simple measurements and are presented above. Only simple data are presented for BD3A since overlapping measurements were not made on this core.

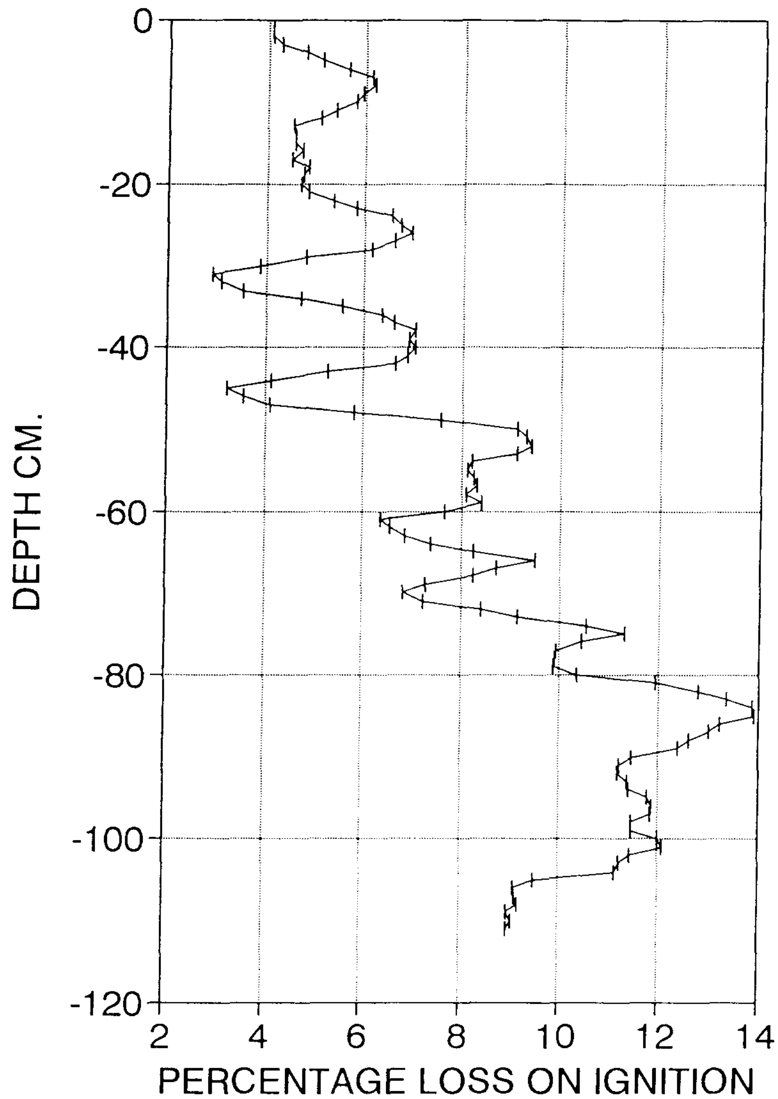
sediments is relatively undisturbed.

4.2.2.3 Organic Matter Content (Loss on Ignition)

The organic content of the core was determined by loss on ignition (Belcher and Ingram, 1950). Samples of about 5 cm^3 were taken from the core at 1 cm intervals. First the samples were oven dried for 24 hours at 105°C. They were then cooled in a desiccator and weighed, after which they were ignited in a muffle furnace at 550°C for an hour. To determine the loss on ignition the samples were reweighed after they had cooled.

Variation in organic content of the core, as measured by loss on ignition shows considerable small scale variability. Whilst this may be real, at the scale of the present study it is uninterpretable and is therefore treated as noise. Hence, the data are presented as a five point moving average (Figure 4.10).

Figure 4.10 Five Point Moving Average of Loss on Ignition Data From Core BD4A



Chapter V Interpretation of Non-Palynological Evidence of Environmental Change

In this chapter the results of non-palynological analyses of core BD4A presented in Chapter IV are discussed. Interpretation of this data is necessarily speculative. Each aspect of the analysis provides part of an interpretative jigsaw. Therefore in the discussion below supporting lines of evidence are drawn together in order to infer the nature of Neoglacial environmental change in the catchment.

5.1 Discussion of Non-Palynological Evidence

Of the various non-palynological techniques used to describe the Blowdown Lake core, the identification of conifer needle macrofossils allows the most definitive interpretation. Unlike pollen, macrofossils are poorly dispersed: "the majority of plant macrofossils found in a sediment body are found locally" (Lowe and Walker, 1984, p.177). The results are therefore highly site specific, in this study the presence of conifer needles in the lake sediments, implies that conifers grew around the lake during the period of deposition of the sediments. The major feature of Figure 4.4 is the shift from higher concentrations of needles in the sediment below 60 cm. depth, to lower values above this horizon. Because of the relatively low numbers of needles of each species found in the core (see section 4.2.1.1) the changes in individual species, are regarded as less significant than the general decline noted in all species. Currently the lake lies at the border of the parkland and krumholz zones of tree growth. Although there are trees close to the lake, needles are sparse in the upper sediments relative to the lower horizons. Therefore the needle evidence is interpreted as indicative of higher treeline in the period represented by the lower half of the core. The dramatically greater needle concentrations in the lower half of the core suggests dense coniferous growth around the lake, perhaps approaching closed canopy forest. It is uncertain whether the minimum in needle concentrations between 80 and 90 cm depth is due to real change in forest conditions, or whether it is simply a reflection of variation in preservation of needles.

Care must be taken in interpreting changes in needle concentration directly as shifts in treeline. As treeline retreats remobilisation of needles from the former forest floor, or preserved in soil humus may maintain needle input to the lake. There may be therefore, a time lag between vegetation change at the site and reduced input of needles to the lake.

The evidence derived from the extent of charring of the needles (Figure 4.5), largely supports the suggestion of higher treeline in the period represented by the lower half of the core. The assumption is made that the charring of the needles is due to the incidence of wildfire within the lake catchment. Assessments of fire history are typically made using counts of microscopic charcoal (Patterson *et al.*, 1987). Microscopic charcoal (soot) is well dispersed by wind during wildfire. Therefore such counts presumably identify a regional fire history. Assessment of charred needles should identify a local signal, in the same way that the needles provide a picture of local vegetation.

Charred needles are present in the lake sediments at all horizons below 55 cm. Warmer summer temperatures during this period would enhance the likelihood of wildfire. However the observed pattern may also be a function of vegetation type at the site, because fire may not spread easily into the subalpine parkland zone where trees are more dispersed. There is a significant peak of charred needles in the sediment at 75 - 80 cm. This peak may represent a single fire. The lower percentages of charred needle fragments at other depths most likely represent a background level of charred fragments from previous fires, washed in from soil storage.

The loss on ignition values for core BD4A (Figure 4.10) are interpreted a representing the balance of organic and mineral material washed into the lake, since autochthonous production is minimal. Autochthonous material in lake sediments may include chemical precipitates, and biogenic material, but for freshwater lakes, in areas of acid lithology, such as Blowdown Lake, the former source is negligible. Production of both mineral and organic autochthonous material is related to the productivity level of the lake. The oligotrophic lakes of highland areas are typically unproductive. In addition steep slopes and high precipitation values favour erosive input of catchment material. Pennington and Lishman (1971) state:

> `We consider that it can no longer be doubted that, in lakes such as those of the English Lake District and the Scottish Highlands, the sediment column is derived from a series of soils originating in the catchment area. These lakes lie on the course of vigorous streams in areas of high relief;....'

Blowdown Lake is an oligotrophic lake, in a high relief area, with two steep inflowing streams. It seems reasonable therefore, to assume that the ratio of organic to inorganic material in the lake sediment is a function of the balance of organic and mineral material washed into the lake. Souch (1984) calculated C:N ratios for lake sediment in another alpine lake in the southern Coast mountains, Gallie Pond. The results suggested that allochthonous sedimentation was dominant. Measurements of loss on ignition therefore provide a crude index of vegetational productivity within the catchment. Obviously the rate of erosion of mineral soil will also affect the measure. However it is assumed that the primary control on erosion of the mineral soils will be vegetation cover. Therefore the effect of changing mineral inputs is merely to amplify the vegetational signal.

The loss on ignition values in Figure 4.10 exhibit considerable structure. Three zones may be identified, From the base of the core to 85 cm, LOI values are relatively high (10-12%). Between 85 cm and 55 cm there is a gradual decline to values of around 8 percent. Above 55 cm the organic content fluctuates between 4 and 7 percent. The minima at 30 cm, 45 cm, 69 cm, and 78 cm are associated with sand layers in the core. They therefore represent single events rather than major climatically forced changes in sedimentation. Based on the assumptions outlined above, these results are interpreted in terms of vegetational change within the catchment.

High organic content in the sediment from the base to 85 cm suggests more productive catchment vegetation than the present during the period represented. The decline in LOI values between 85 and 55 cm is consistent with treeline retreat in this period. Above 55 cm the LOI values fluctuate in a range close to modern values. With the exception of the minima discussed above these fluctuations are assumed to represent minor changes in sedimentation, associated with smaller vegetational changes, or changing sediment sources. The LOI evidence therefore suggests a pattern of Neoglacial catchment change similar to that interpreted from the macrofossil evidence above. Interestingly LOI values start to decline at the 85 cm level whereas the major decline in conifer needle concentrations occurs around 55 cm. It may be that a climatic decline led to break up of the forest canopy at the site, and progressively increasing erosion as treeline retreated. However it is hard to reconcile decreasing organic inputs to the lake, while concentrations of conifer needles increase sharply during this period. A possible explanation is that treeline retreated to below the lake during the period represented by the sediments at 85 cm depth. Increased erosion of mineral soils in the catchment would account for declining LOI values. In this case the persistence of needles in the sediments after this period may be associated with erosion and redeposition of needles preserved in the humic layer of the forest soils. Alternatively treeline may not have retreated away from the lake until the period identified by the major decline in needle concentrations. In this case the lower LOI values may be due to increased erosion of mineral material just from the upper slopes where treeline retreat had already occurred. Similarly increased duration of snow cover at high elevations might limit development of ground cover vegetation, increasing supply of mineral sediment, and decreasing organic inputs to the lake. It appears therefore that interpretation of macrofossil evidence in terms of changing treelines is not straightforward.

Of the three principal features of core stratigraphy noted, the colour change associated with changing organic content and the visible needles are quantified and interpreted by the techniques discussed above. The remaining features of note are the sand layers.

The distinct sand layers apparent in the stratigraphy of core BD4A (Figure 4.6) are interpreted as representative of episodic high magnitude events, resulting in input of relatively coarse sediment to the lake. Examination of the potential sediment sources for the lake suggests three possible origins for the sand layers. They may be representative of debris flow events generated on the talus above the lake; they may be deposited by major floods, perhaps associated with spring snowmelt, or rain on snow events; or they may be deposited by large avalanches carrying material into the lake.

Debris flow tracks are evident on the talus although currently, none approach the lake shore. The base of the talus is ringed by large boulders which could potentially limit access to the lake for small debris flows.

As noted earlier, in the summer of 1993 there was extensive evidence of avalanching during the previous winter. Tilted trees and debris suggest that avalanches commonly reach the edge of the lake. The upper half of the avalanche slopes are marked by unvegetated areas, although there is no evidence of soil erosion in the avalanche runout zone. Occasional full depth avalanches may cause erosion of soil from the slopes may therefore be a significant sediment source.

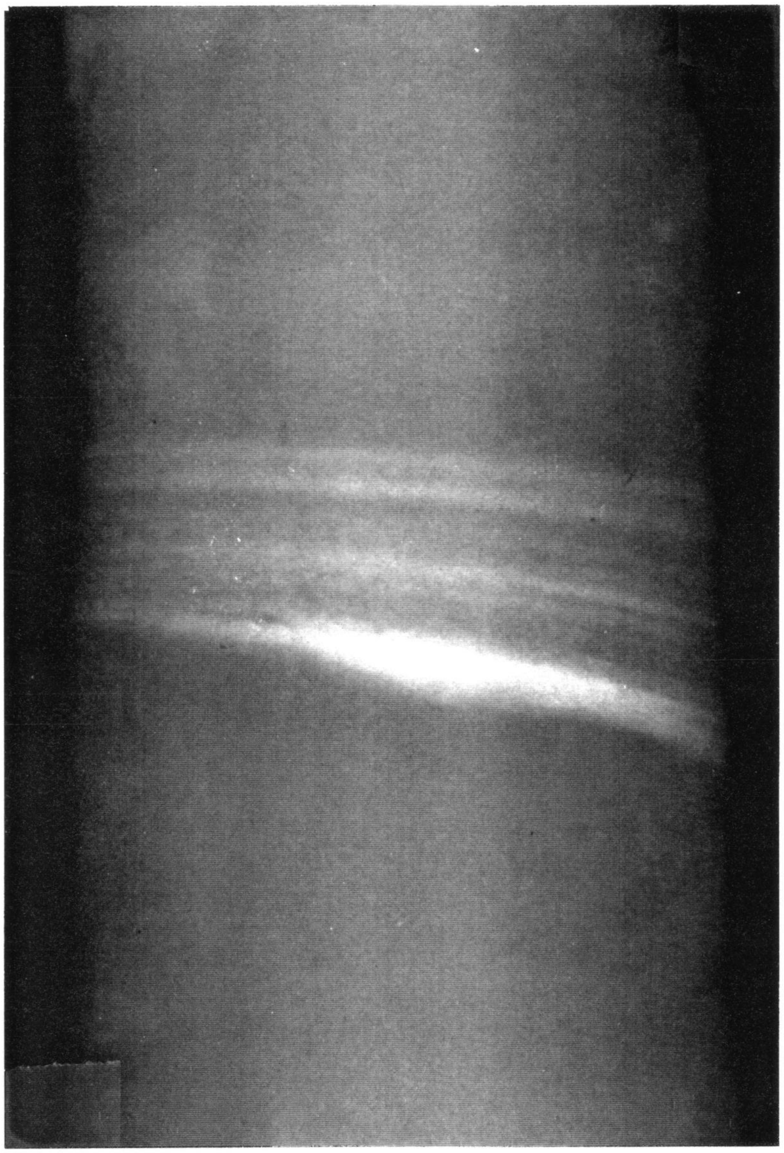
From the available evidence it is not possible to identify the process by which the sand layers were transported to the lake. However variability in the character of the layers may provide some clues.

The coarse upper sand layers associated with dropstones presumably result from higher energy events than the finer sand layers below. The presence of dropstones suggests that the sediment was initially deposited on winter lake ice, and then settled subsequently when the ice melted. This may suggest that avalanching is the source of the sediment. It does not however exclude other sources. The fact that there are no dropstones in the lower sand layers suggests that the events which deposited these layers were much lower energy, or else that they took place when there was no lake ice. This hints at higher temperatures during this period, which is in line with the evidence of lower treeline.

Careful scrutiny of the x-ray plates of the sand layers in the lower half of the core, reveals two cases of laminations occurring above the initial inflow event. In each case there are 10 - 20 fine laminations each of about a millimetre in thickness (Figure 5.1). It is uncertain whether these laminations represent synoptic scale events, or whether they are annual rhythmites. Given the slow sedimentation rates found generally in the core, the latter appears more likely. This suggests that some process associated with the original event resulted in increased sediment yield to the lake during one part of the year for 10 - 20 years after the event. A likely scenario is that the initial event stripped vegetation from the basin, allowing increased surface erosion during summer storms. 10-20 years seems a reasonable period for vegetation regrowth and slope stabilisation. The most likely event to strip vegetation in the catchment is forest fire. However evidence from charred needles doesn't suggest a peak in fire activity associated with the most apparent laminated horizon (66 - 68 cm), although charred fragments of wood were recovered from the inwash layer. It is also possible that a major mass movement such as a debris flow, or erosive slush avalanche could expose bare soil and produce a similar effect.

Interpretation of mineral magnetic stratigraphy in terms of environmental variables is complex. Sources of magnetic material to the lake may include both

Figure 5.1 X-Ray Photograph of Laminae Associated With the Sand Layer at 66-68 cm Depth in Core BD4A. (Lighter shades represent denser material)



allogenic sources (eroded sediment, aeolian deposition), and authigenic material (chemically or bacterially produced). However, Thompson and Oldfield (1986, p102) state:

> "the studies completed so far have led us to regard magnetic minerals in lake sediments as overwhelmingly allogenic except where there is positive evidence to the contrary"

Considering controls on delivery of allogenic material to the basin, they identify both vegetation cover, and hydroclimate, as important variables. The implication therefore is that magnetic minerals are eroded and transported to the lake more easily under wetter, or more poorly vegetated conditions. A frequently cited study by Hallam *et al.* (1973) clearly identifies parallelism between change in magnetic susceptibility and pollen evidence of vegetational change in lake sediments from High Furlong, Lancashire. Oldfield and Robinson (1985) interpret this evidence as representative of cool phases with low vegetation cover and soil instability, causing input of unweathered magnetic minerals; and warmer vegetated phases with soil development and lower erosive input of magnetic minerals. Similarly Thompson and Oldfield (1986) report that Saturation Isothermal Remnant Magnetisation (SIRM) values of lake sediments (another measure of magnetic mineral concentration) closely parallel changes in Na, K, and Mg in the core. Mackereth (1966) suggested that these soluble elements are good chemical indicators of soil erosion. These results are further evidence that magnetic mineral concentrations in lake sediments are closely related to erosion rates within the catchment.

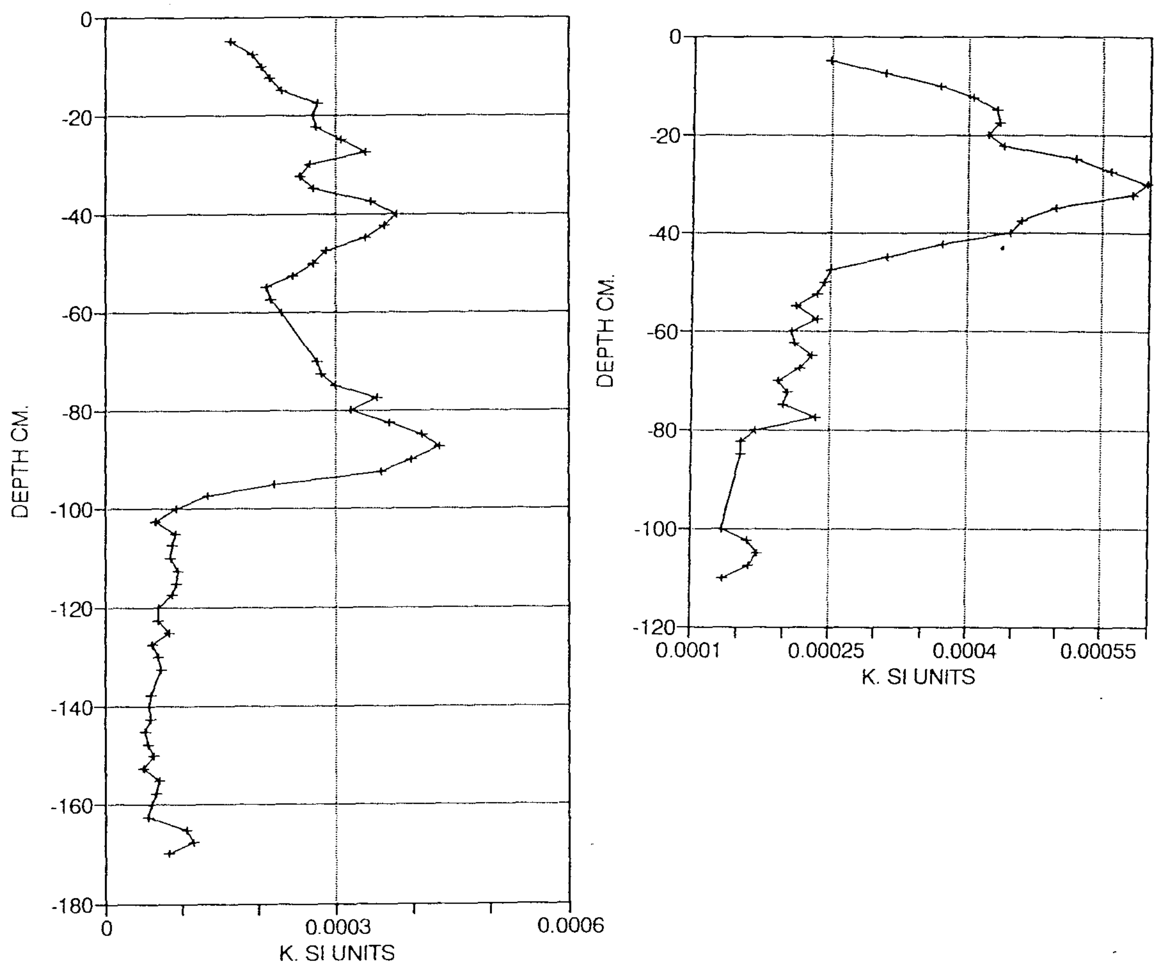
Figure 4.9 compares k values for two cores from Blowdown Lake. Core BD4A is from the deepest part of the lake whilst BD3A is a core taken from closer to the shore. The k values for BD3A are consistently higher than those for the main core from the centre of the lake. This supports the assumption that allogenic sedimentation dominates in the lake. Thompson and Oldfield (1986) note that commonly, nearshore sediments have higher magnetic susceptibility than sediments from the centre of the lake, arguing that this supports an allogenic origin for magnetic minerals within the sediment.

A minor peak in susceptibility at 77.5 cm depth corresponds with the major peak in input of charred needles. It seems likely therefore that this magnetic peak is associated with post fire input of fire generated secondary magnetic minerals. Studies by Rummery (Rummery *et al.*, 1979; Rummery, 1983) have identified magnetic susceptibility peaks in lake sediment associated with fire episodes. These are in part due to the formation of secondary magnetic minerals by heating during the fire. However it is likely that clearance of vegetation by the fire, and increased surface wash and erosion is also a factor.

The magnetic profile of the Blowdown Lake core BD4A exhibits the same general zonation as the LOI curve (Figure 4.9. From the base to 80 cm depth k values are low, corresponding with high organic content in the sediment. From 80 cm to 50 cm there is a gradual increase in k, followed by a major peak in k at 30 cm depth. Subsequently k values decline again. The period represented by sediments above 50cm corresponds to the most minerogenic sediment in the core. The parallelism between the LOI curve and changing k values supports the assumption that change in k is associated with change in erosion in the basin (since authigenic mineral production is primarily non-magnetic silica). Comparison of magnetic data from Blowdown Lake with the magnetic profile of the Upper Kidney Lake core (Figure 5.2) is of interest. Despite the fact that these profiles are from different lakes, they exhibit the same basic form, with a rapid transition from low to higher k values. This suggests that some forcing factor external to the lakes is responsible for the observed change in mineral magnetic content of the lake sediments. This factor is assumed to be climate. Change in climate affects sedimentation in the lake by directly impacting erosion rates, and by inducing vegetational change, which affects runoff and erosion in the catchment. Oldfield and

Figure 5.2 Comparison of Magnetic Stratigraphy of Cores from Blowdown Lake (BD4A) and Upper Kidney Lake (KL1A)

KL1A **BD4A**



MAGNETIC SUSCEPTIBILITY TRACE
5 CM MOVING AVERAGE

MAGNETIC SUSCEPTIBILITY TRACE
5 CM MOVING AVERAGE

Robinson (1985) point out that the usefulness of mineral magnetic parameters as palaeoclimatic indices is a function of climatic control on sedimentological processes. In this case this control is further moderated by vegetational control on erosion.

The magnetic evidence from the core supports the general pattern of change derived from other evidence. Low values at the base of the core imply stable conditions. A transitional phase is marked by increasing erosion, and the period represented by the upper 50 cm of the core has experienced apparently much increased erosion. It is however unclear why K values decline above the 30 cm level. LOI values remain low in all sediments above the 50 cm level. Three hypotheses to explain the phenomena are advanced below.

1) The Blowdown Lake catchment lies within the plume of the Bridge River tephra (Mathewes and Westgate, 1980) however the tephra is not visible in the core stratigraphy. The tephra dates from 2350 BP, given the radiocarbon date from close to the base of core BD4A of 4140 BP the absence of the Bridge River tephra is unusual. It is suggested that the peak in K values may represent a 'smeared' signal due to inwash of tephra from the slopes of the catchment, since inputs of volcanic ash to lakes cause peaks in magnetic susceptibility of the sediments (Thompson and Oldfield, 1986).

2) In some catchments the mineral magnetic component of sediments is not a result of primary magnetic minerals, but due to the formation of secondary magnetic minerals through chemical processes in soil (Thompson and Oldfield, 1986). If this were the case in the Blowdown catchment, the peak in k may represent inwash of secondary magnetic minerals, formed in the forest soils during the period of higher treeline. The subsequent decline in k values would then represent exhaustion of the supply of magnetic minerals from the former forest soil. Under this scenario it is necessary to invoke a time of much increased soil erosion (probably climatically induced) in the period represented by sediments above 50 cm. This would explain the major peak in k values occurring well after conifer needle macrofossil evidence suggest treeline has

retreated.

3) k is a measure of volume susceptibility. The measurements were taken on wet cores. If the water content in the upper portion of the core were higher, this might artificially depress k by reducing the concentration of magnetic minerals. If this were the case measurements of k on dried sediment would be expected to show consistently high k values above 50 cm.

Evidence available in the current study is not sufficient to distinguish between the alternatives outlined above. They are presented as possible explanations. More detailed mineralogical investigation might reveal an interesting magnetic record of hillslope erosion.

5.2 Summary

Integration of the evidence from the various proxy data sources discussed above produces a coherent picture of environmental change within the catchment. The period represented by sediments from the base of the core to 85 cm was apparently marked by much denser arboreal vegetation around the lake. Presumably this higher treeline was associated with higher summer temperatures. During this period sedimentation in the lake was relatively organic, implying stable vegetated slopes and high productivity within the catchment. Magnetic susceptibility measurements suggest relatively low inputs of mineral soil to the lake. Forest fire was also apparently a feature of this ecosystem.

The period represented by sediments at depths of 85 cm to 60 cm was one of transition. The physical sedimentology suggests increasing erosion within the catchment. Presumably this was associated with vegetation disturbance, resulting from a climatic decline. Inputs of needles to the lake remain high during this period.

In the period represented by sediments above 50 cm conditions at the site were apparently similar to the present. The lake is above treeline and relatively high catchment erosion rates produce dominantly minerogenic sediment in the lake. Under these cooler climate conditions fire does not appear to be an important factor in the landscape.

Chapter VI Palynological Evidence of Environmental Change

6.1 Surface Pollen Samples

6.1.1 Percentage Data

Figure 6.1 presents the results of pollen analysis of the surficial sediment samples, this percentage diagram exhibits little elevational zonation of pollen deposition which is perhaps unsurprising, given the difficulties in interpretation of percentage pollen data in mountainous areas (outlined in 2.3.1). The two surface sample at 2057 and 2060 m are from the slopes to the north of the lake, and the ridge to the south respectively. They do not fall on the linear transect from the forest to the lake, from which the other samples were taken and therefore are not strictly comparable. They have been included in the data set because of the lack of suitable sampling sites above the lake, on the line of the transect. It is however important to note this potential source of error in consideration of the results.

The lower reaches of the Blowdown valley were logged in 1981, and subsequently replanted in 1988. Recent pollen deposition may therefore not be strictly representative of normal vegetation in the area. However the logging was 3 km from the lake at its closest point. The effect on pollen deposition would therefore be predominantly on the extra-local component of the pollen rain. Surface samples were from hollows and the lake (which is small), therefore local pollen should dominate the pollen spectra. 3 km (elevation change 350 m) is further than treeline might be expected to vary during the Neoglacial. Therefore if it were the case that extra-local pollen deposition were important at the site, it would not be possible to detect treeline shifts anyway. It is therefore reasonable to attempt to interpret downcore variability on

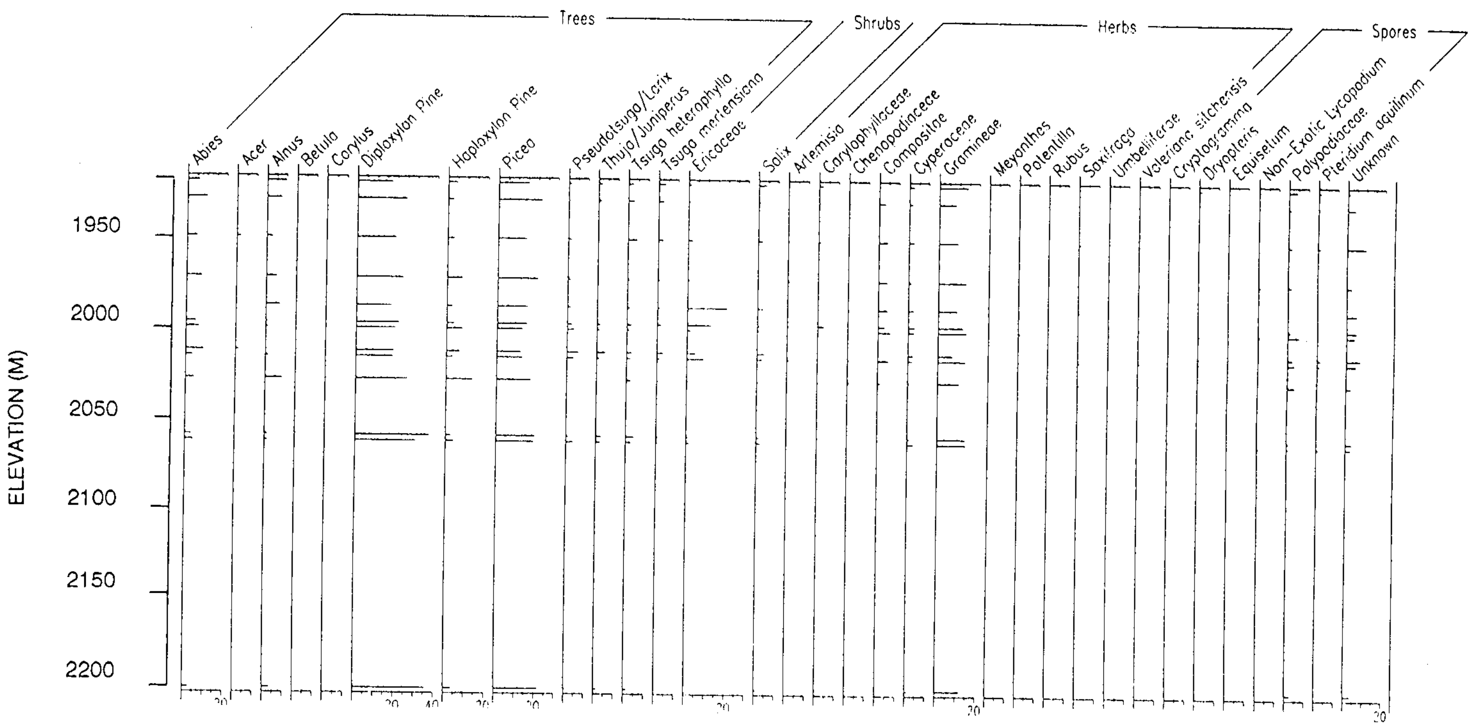


Figure 6.1 Blowdown Lake Catchment Surface Samples - Percentage Pollen Data

the basis of the surface pollen samples obtained. At sites between 1990 and 2020 m elevation high values of *Ericaceae* pollen are recorded. Associated with these are relatively low values of the principal coniferous pollen types. The high *Ericaceae* values may simply be a function of very local pollen production, reflecting microtopographic control on plant distribution. However, the principal ericaceous shrubs in the area are *Cassiope mertensiana* and *Phyllodoce glanduliflora*. These are most abundant in a relatively well defined region of well drained areas below the lake. The explanation for the pattern of pollen deposition may therefore lie in a phenomena identified by Frenzel (1969). He noted that the ratio of non-arboreal pollen to arboreal pollen was highest at treeline. Above treeline, low local production is swamped by up-valley transport of tree pollen, whereas in the meadows at treeline, local production is enough to depress the percentage of arboreal pollen. The data from the Blowdown catchment may therefore be indicative of depression of percentage values of conifer pollen by high local production of ericaceous pollen at treeline. No surface samples were obtained from between 2060 and 2200m. However, the similarity between the pollen spectra for each of these elevations probably reflects the continuing importance of up-valley transport of arboreal pollen. The percentage pollen data from the surface samples, therefore offer little prospect of quantifying shifts in treeline elevation because of "swamping" (Kearney, 1983) of the signal by lowland pollen.

6.1.2 Pollen Ratios

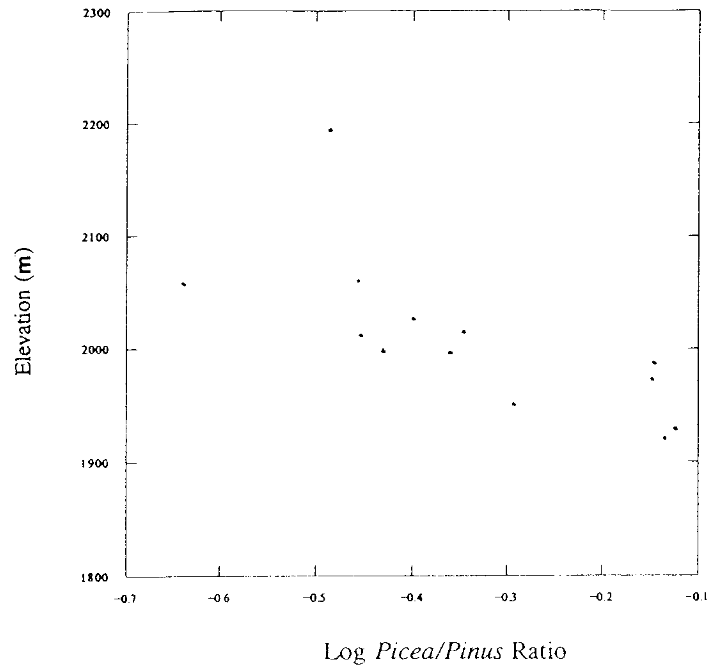
Pollen ratios were derived in an attempt to quantify the pollen deposition/elevation relation. Principal treeline species are *Picea engelmannii* and *Abies lasiocarpa*. *Abies amabilis*, *Pseudotsuga menzeseii*, *Pinus contorta*, and *Tsuga heterophylla* grow in the montane forest. The requirement of the pollen ratio technique is that one pollen type be well dispersed from lower elevations, whilst the other is locally deposited at treeline.

Therefore the following ratios were calculated: *Picea/Pinus, Picea/Pseudotsuga, Abies/Pinus,* and *Abies/Pseudotsuga.* Ratios involving *Tsuga* were not calculated because of the low representation of the species in the pollen spectra. Of the ratios which were calculated, those including *Pseudotsuga* showed no clear relation when plotted against elevation. Theoretically the decline in pollen deposition away from the source should be exponential (Maher, 1963). Therefore a linear relation is expected between elevation and the log of the pollen ratio.

Figure 6.2 illustrates this relation for the *Abies/Pinus* and Picea/*Pinus* ratios. The *Abies/Pinus* plot shows considerable scatter. The *Picea/Pinus* plot exhibits less scatter but the 1996 m surface sample is an outlier on both plots. Because of the lack of data points between 2060 and 2200 m the significance of this cannot be determined. It was decided to disregard this point in further analysis, and limit calibration on the basis of the *Picea/Pinus* ratio to the 1920 to 2060 m elevational range. The 1987 m data point on the *Abies/Pinus* plot is also clearly an outlier and was eliminated. Linear regression of the two data sets produced the best fit lines presented in Figure 6.3. The r^2 value of the *Picea/Pinus* regression is slightly better than that for *Abies/Pinus*. The standard error of the estimate for the former regression is just over half that for the latter. Therefore the *Picea/Pinus* relation was selected to calibrate the downcore pollen record. This calibration is over a more limited elevational range than that for the *Abies/Picea* ratio, but the potential error is less. This is an important consideration given the relatively small fluctuations in treeline expected during the Neoglacial.

It is important to note that the *Picea/Pinus* ratio for the surficial sediments of Blowdown Lake (sample at 2026 m) falls close to the line fitted through the points. Removal of the lake sample from the data set does not significantly alter the best fit line. It is therefore reasonable to assume that the hollows from which the surface samples were collected have similar pollen recruitment characteristics to the lake. Use of the surface data to interpret downcore pollen variability is therefore a valid exercise.

Figure 6.2 Plots of Log Pollen Ratios From Blowdown Lake Catchment Surface Samples Against Elevation



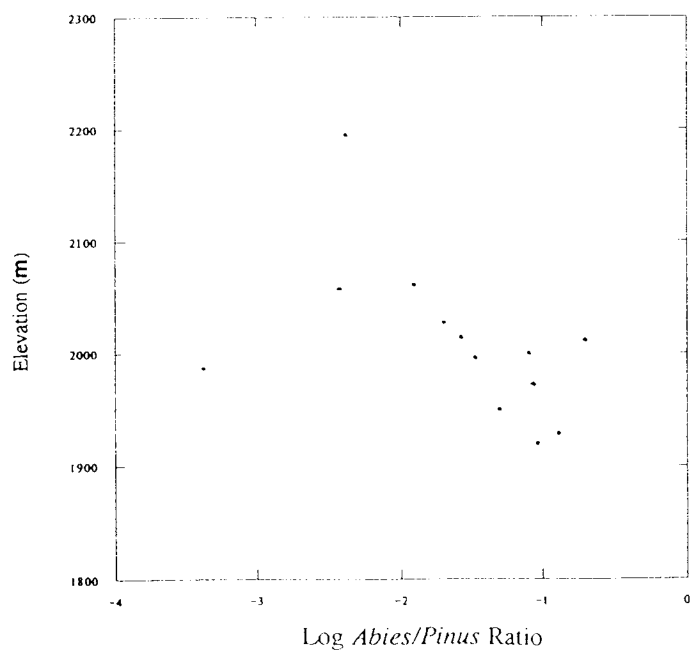
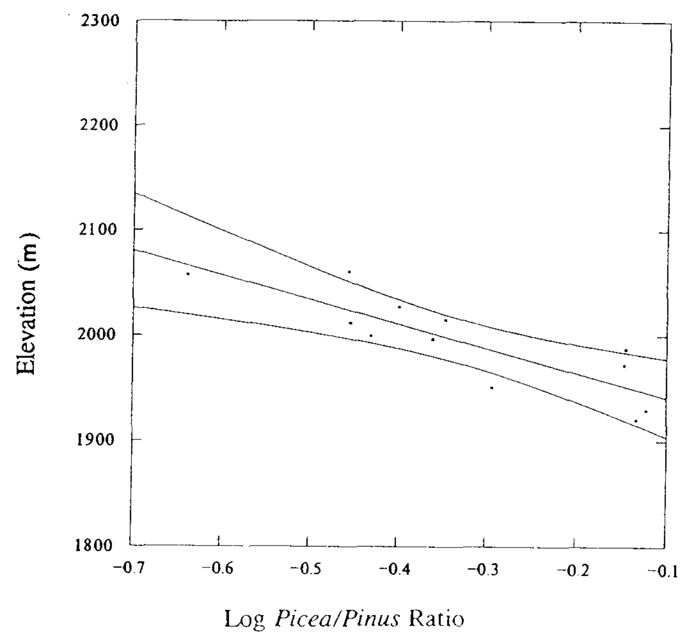


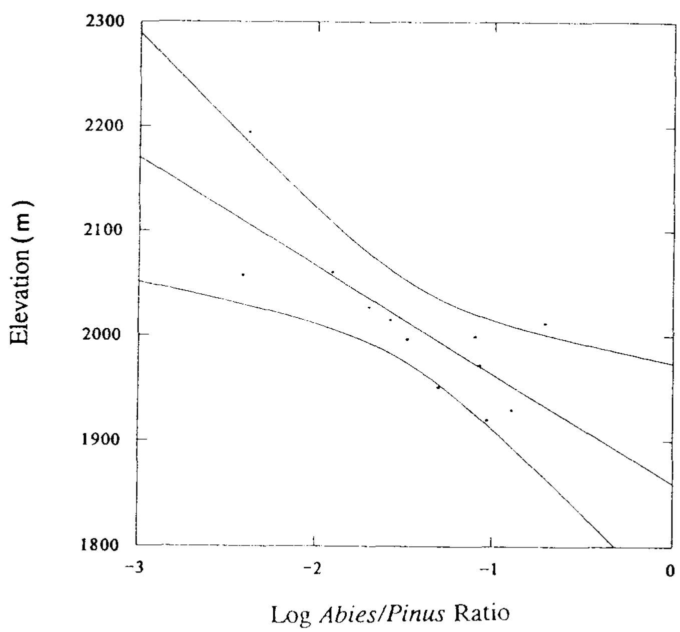
Figure 6.3 Regressions of Log Pollen Ratios from Blowdown Lake Catchment Surface Samples Against Elevation



y = -232.727x + 1917.704

r^2 = .71

Std. Err. Est. = 25.432



y = -103.579x + 1859.561

r^2 = .621

Std. Err. Est. = 47.287

6.2 Blowdown Lake Pollen Analysis

6.2.1 Percentage Data

Figure 6.4 is the percentage pollen and spore diagram of the Blowdown lake data. Three local pollen zones (BD. 1 to BD. 3) have been defined by visual inspection. The three zones are characterised below.

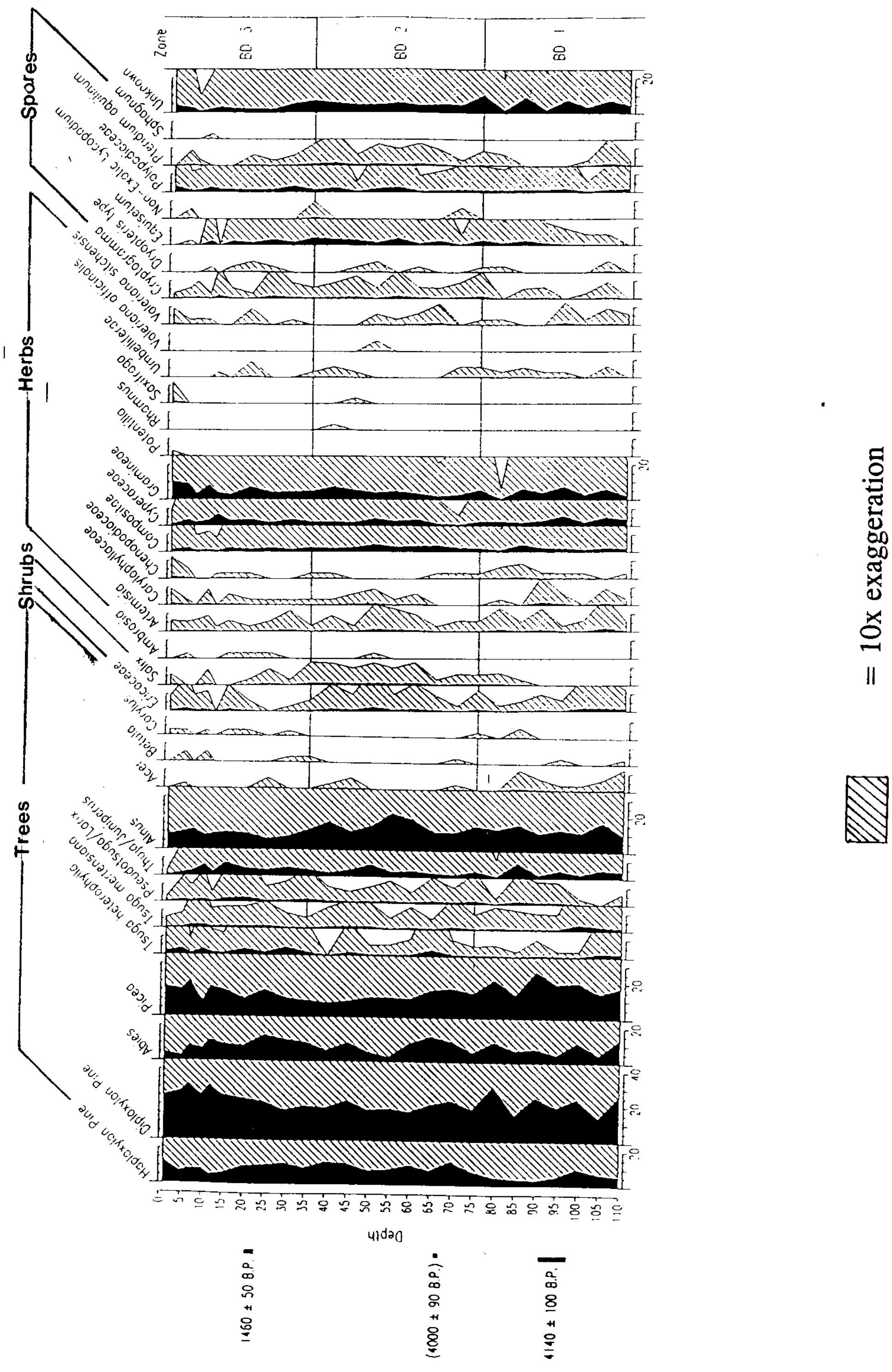
6.2.1.1 Pollen Zonation

BD.1 This zone is marked by high values of coniferous pollen types. Diploxylon Pine and *Picea* fluctuate at close to 20 % whilst *Abies* comprises around 10% of pollen and spores. *Acer* is represented by a continuous curve throughout most of the zone. Haploxylon pine (*P. monticola* or *P. albicaulis*) values are 5-10% at the base of the zone but decline to a minimum of 3% in the upper half of the zone. The upper boundary of the zone is marked by declines in the *Pinus* and *Picea* curves.

BD.2 Zone 2 is characterised by lower values of coniferous pollen types. *Picea* pollen declines from the base of the zone reaching a minimum around 8 %. *Abies* pollen percentages remain high at the base of the core but decline to less than 5% before recovering in the upper part of the zone. Maxima of *Alnus, Ericaceae, Salix* and *Pteridium* pollen occur in this zone. The upper zone boundary is marked by decline in pollen percentages for *Alnus*, *Ericaceae*, and *Salix*.

BD. 3 Zone 3 is marked by a return to higher values for coniferous pollen types. *Pinus* and *Picea* increase, although *Abies* declines from initially high values. *Tsuga heterophylla* reaches its highest values in this zone at around 5%.

Figure 6.4 Blowdown Lake Percentage Pollen Diagram



6.2.1.2 Interpretation of Percentage Pollen Data

As noted in Chapter II interpretation of percentage pollen diagrams from mountainous areas is complex. However in the light of the evidence from non-palynological analyses of the core, and the data from local surface samples, the following interpretation is suggested.

Zone 1 spans 110 - 75 cm in core BD4A and therefore approximates the period of higher treeline suggested by the non palynological evidence (depths 110 - 85 cm). High percentage representation of coniferous pollen types, in particular *Picea* support this interpretation. Five to ten percent *Abies* pollen, and fifteen to twenty percent *Picea* are similar values to those recorded in the lowest surface pollen samples from dense subalpine forest.

Zone 2 spans 75 cm to 35 cm depth which covers all of the 'transitional' phase identified from sedimentary evidence (85 - 60 cm). Low percentages of *Picea* throughout the period suggest a retreat in treeline. Lower values of *Abies* support this interpretation, although the pattern is less clear. *Salix, Ericaceae,* and *Alnus* are common species of the sub-alpine parkland zone at treeline. If treeline retreated, causing a transition in vegetation around the lake from sub-alpine forest to parkland, an increase in these species would be expected. It has already been noted (section 5.3.1) that currently, high *Ericaceae* pollen percentages depress representation of coniferous pollen types just above treeline at the site. Peak values of *Ericaceae, Alnus,* and *Salix* in zone two therefore provide further indication of treeline retreat during this period. The peak in *Pteridium* spore percentages during this time period is difficult to interpret. Currently *Pteridium* does not grow at the site. The species is an early coloniser of burnt areas, and therefore indicates that fire is an important component of local ecology. If the *Pteridium* spores were of local provenance it would suggest increasing fire frequency during this period. Given the fact that treeline was apparently retreating, this is unlikely. Treeline retreat is indicative of cooler temperatures, and reduces the fuel supply for wildfire. In addition, the evidence from charred conifer needles (Figure 4.5, Section 5.1) suggests decreased fire frequency above 50 cm. The *Pteridium* spores may therefore be due to long distance transport from lower elevations. However increased fire frequency during a period of increasing coolness is still unlikely. Despite this, persistence of *Pteridium* spores into zone 3 suggests that long distance transport of *Pteridium* spores occurs, since *Pteridium* does not currently grow at the site. An alternative explanation of the peak in percentages of *Pteridium* spores, is that it represents secondary remobilisation of the spores. Birks and Birks (1980) note that spores are rich in sporopollenin, and are therefore differentially preserved in soils. It has already been suggested that rates of soil erosion may have increased during this period (Section 5.1). The peak in representation of *Pteridium* spores may therefore represent inwash of preserved spores from eroding forest soils.

Zone 3 covers the depths from 35 cm to the surface. The nature of the percentage pollen diagram during this period is similar to that of zone 1 with high *Picea* and *Abies* pollen percentages. However non-palynological evidence suggests that rather than a return to high treeline, this period is marked by a continued cooling of climatic conditions (5.2). The zone 3 pollen spectra are therefore interpreted as representing sub-alpine/alpine conditions similar to those at the lake today. Low local pollen production by krumholz, grasses, and alpine herbs is therefore 'swamped' (Kearney, 1983) by up valley transport of lowland coniferous pollen. Surface pollen samples from the lake contain high percentages of conifer pollen, despite being above treeline.

Tsuga heterophylla pollen reaches its highest percentage in zone 3. This species is not present at the site today, but grows at lower elevations. *Tsuga heterophylla* is a component of the wet forests of the coastal zone, and is uncommon in dry interior forests. The site lies on the coast-interior ecotone. The increase in *Tsuga heterophylla* pollen is thus interpreted as indicating a shift to wetter conditions in the lower valley. The factor controlling this distributional shift may in fact be fire frequency. *Tsuga heterophylla* has thin bark and hence low resistance to fire and is therefore sensitive to fire frequency (Minore, 1979). However reduced fire frequency in the lower valleys is most likely to be associated with wetter and perhaps cooler conditions.

6.2.2 Pollen Concentration Data

Figure 6.5 illustrates pollen concentration data for selected taxa. Given the apparently episodic nature of sedimentation in the lake no attempt has been made to construct an influx diagram. It is therefore important to bear in mind that the concentration data presented are an integration of pollen production, and rates of sedimentation. The pattern of pollen concentration change is similar to the structure of the percentage pollen diagram. However the increase in coniferous pollen in zone three is less apparent. This may support the assertion that this apparent increase is an artefact of the percentage pollen diagram. Alternatively it may simply be a reflection of increased sedimentation rates during the period represented by zone 3. The transition from high tree pollen values to lower between zones 1 and 2 is emphasised in the concentration diagram. Again this could be a function of changing sedimentation rates. However it is suggestive of a shift in vegetation from forest, to less productive sub-alpine shrubs and herbs.

Total pollen concentrations are illustrated in Figure 6.6. The general pattern of high concentrations from the base of the core to 75 cm, followed by an extended period of lower concentration, appears to support the interpretation of higher treeline during the period represented by sediments below 75 cm. Owens (1990) suggests that sedimentation rates in the alpine zone of the Coast Mountains are 1-2 orders of

Figure 6.5 Blowdown Lake Pollen Concentration Diagram - Selected Taxa

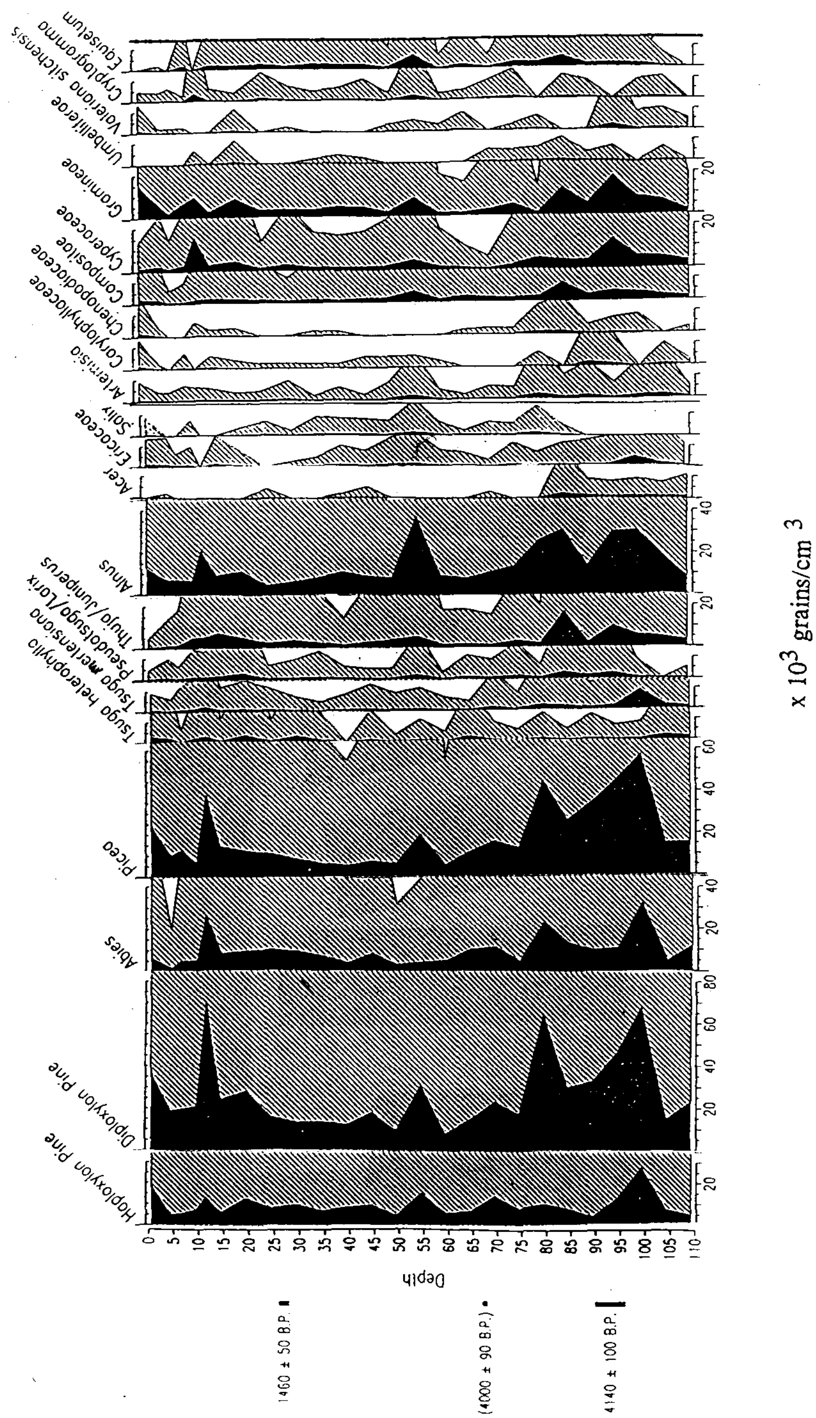
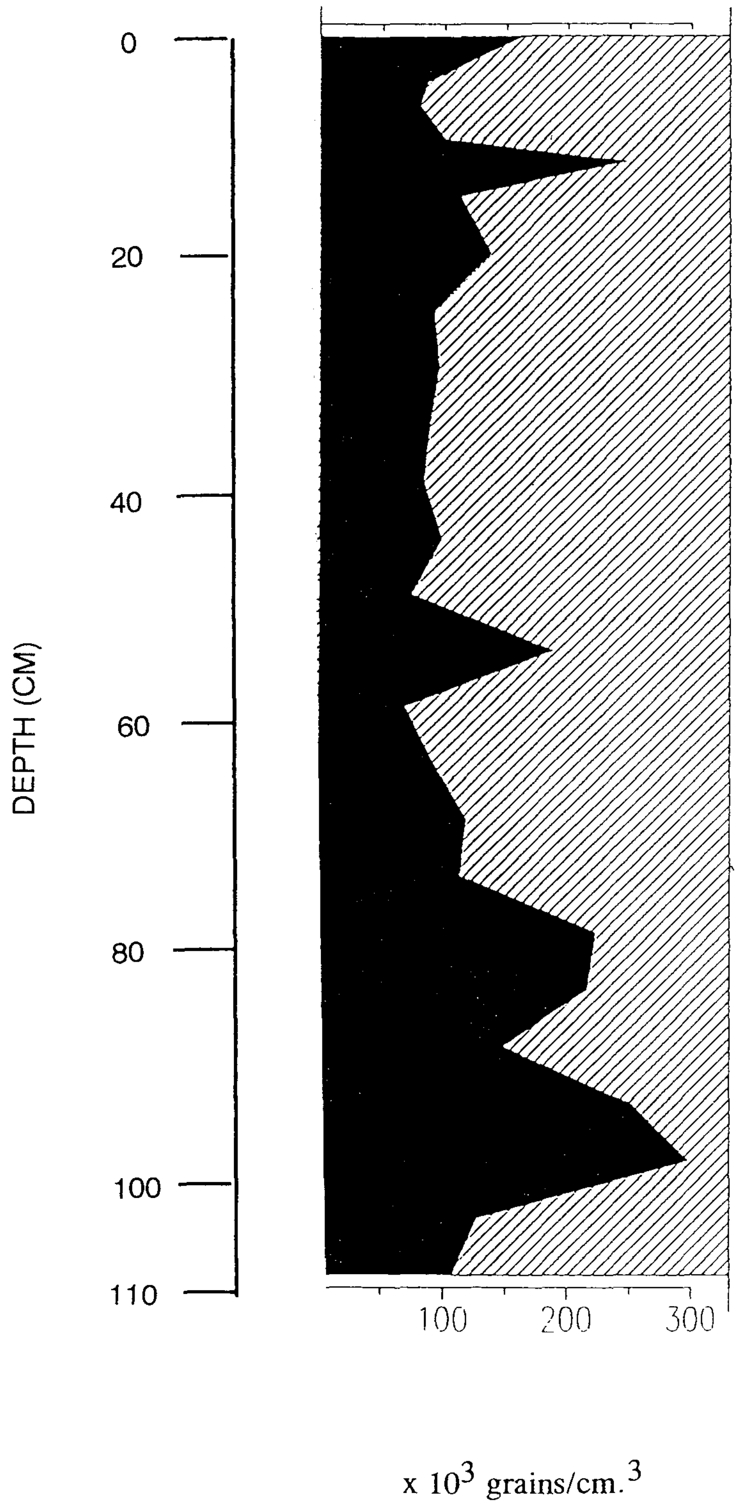


Figure 6.6 Blowdown Lake Total Pollen Concentration



magnitude lower than below treeline. Mathewes (1981) notes that alpine lake sediments may have high pollen concentrations because of low sedimentation rates, even though pollen production is low. On this basis high basal pollen concentrations in the Blowdown lake core would represent lower treeline. However, the depth of sediment deposition since 2350 B.P.[1] at the deepest point of Blowdown Lake, is 2-3 times that of Owens (1990) study sites. Given the macrofossil and sedimentological evidence of higher treeline in the period represented by sediment below 75 cm it is assumed that variation in sedimentation has not been sufficient to mask changes in pollen productivity at this site.

6.2.3 Pollen Ratio Data

Downcore Picea/Pinus ratios calculated from the pollen data are presented in Figure 6.7. Interpretation of this data rests on the assumption that the *Picea* pollen is primarily local, treeline derived, pollen whilst the *Pinus* pollen is a regionally derived, relatively constant value (see Chapter II). Therefore high values of the ratio are indicative of trees close to the lake, hence higher treeline, and *vice versa*. In order to qualify the interpretations offered below the nature of this assumption at Blowdown Lake is briefly examined.

Maher (1963) noted that the pollen ratio technique should only be applied where vegetational communities have been stable and of similar density over time (Section 2.3.4). At Blowdown Lake the species composition of local and regional vegetation, interpreted from the pollen diagram is relatively constant. However there are indications that

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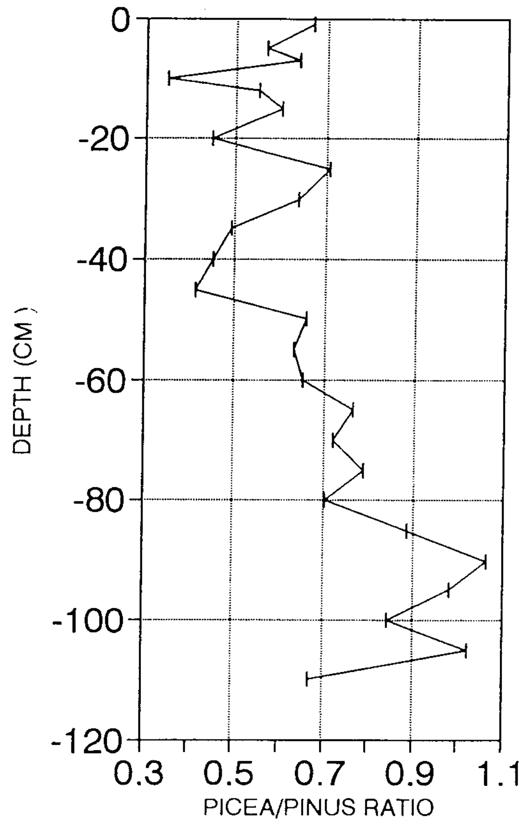
1 Owens presents depths of sediment to the Bridge River Tephra (2350 B.P.) For this comparison linear interpolation between the two acceptable ^{14}C dates on core BD4A was used to estimate the depth of sediment corresponding to 2350 ^{14}C years of deposition. For more detailed discussion of interpolation and dating refer to Chapter VII.

Figure 6.7 *Picea/Pinus* Ratio Variation - Core BD4A



fire may have influenced local ecology in the period represented by sediments below 85 cm, potentially causing local clearings. The possible effect of logging on surface sample representativeness was noted in 6.1.2. The fact that the samples at 1, 5, and 7 cm depth exhibit similar *Picea/Pinus* ratios, supports the assumption that surface pollen spectra have not been dramatically altered by logging at the bottom of the valley. Even at a conservative estimate 7 cm. represents 200 years of sedimentation, extensive logging in the local area has only occurred in the last 20 years. In addition the complex nature of controls on high altitude pollen deposition (Section 2.3.2) may significantly distort the relation between pollen ratios and treeline elevation. However (Kearney,1983) states that pollen ratios have more potential for interpreting elevational shifts in vegetation than percentage pollen data. This is especially the case where parallel lines of evidence substantiate interpretation of the pollen ratio record.

Even after considering possible errors however it is argued that the *Picea/Pinus* ratios should yield at least an approximate record of treeline shifts.

The general pattern of change in the *Picea/Pinus* ratios is high values below 80 cm, a transitional period between 80 and 50 cm, and lower values above 50 cm. There is some structure to the curve in the latter period. Low values between 50 and 35 cm are interrupted by as brief return to higher values between 35cm and 20 cm. Another period of fluctuating lower values succeeds this above 20 cm. The pattern of high treeline, a transitional period, and then lower treeline is the same pattern identified from sedimentological, macrofossil, and pollen percentage evidence. This strong qualitative similarity provides good support for the assumption that the *Picea/Pinus* ratio is a useful proxy measure of changing treeline. In addition the structure of the curve during the period of lower treeline suggests that a record of treeline fluctuation in response to Neoglacial climatic deteriorations may be preserved at the site.

Equation 1, derived from the surface pollen samples (Figure 6.2) was used to calibrate the downcore record in terms of past apparent elevations of the lake.

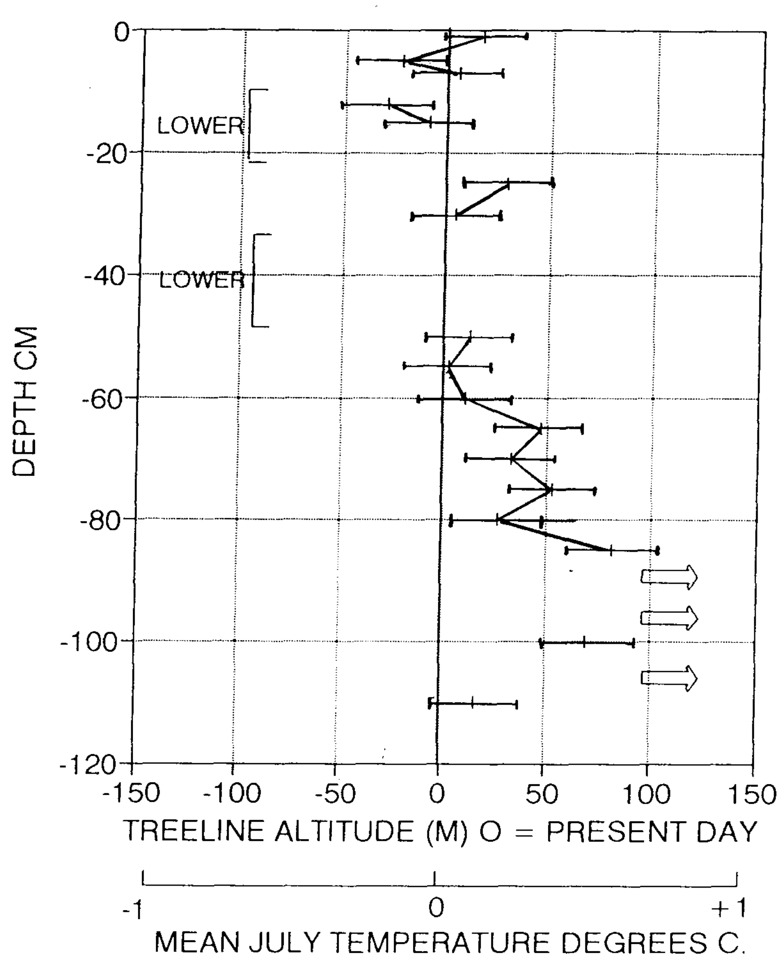
Equation 1. $y = -232.727x + 1917.704$

Since these elevations are determined with respect to treeline, subtracting the current elevation of the lake from this past apparent elevation gives the change in treeline altitude. Calculated change in treeline at the site is plotted in Figure 6.8. Samples in the downcore record which fell outside the calibration range are not plotted. However a qualitative indication of the probable direction of treeline change is given for these points. Also plotted on this figure is a scale representing mean July temperatures. It is assumed that these are the principal control on changes in treeline elevation at the site. The temperature scale is based on a calculated local lapse rate of 0.7°C/100 m (see Appendix 3). The calibrated pollen ratios suggest that treeline was at least 80 m above present in the period represented by sediments from 110 cm to 85 cm. This corresponds to July mean temperatures at least 0.5°C. above the present. Treeline declined to approximately its present altitude whilst the sediments from 85 cm to 50 cm were deposited. Subsequently, treeline has fluctuated around its present level. Two distinct periods of cooling are identified. The magnitude of these cooling events cannot be estimated qualitatively since the pollen ratios fall outside the calibrated range.

6.2.4 Summary of Blowdown Lake Pollen Evidence

The palynological evidence of vegetational change at Blowdown Lake is consistent with the pattern of environmental change deduced from non-palynological evidence. High treeline (below 85 cm in the core) implies warmer summer temperatures than the present. A transition to cooler conditions culminates in two periods of lower treeline

Figure 6.8 Treeline Change at Blowdown Lake During the Neoglacial Period



Error bars = one standard error.

than at present. The percentage pollen data clearly illustrates the difficulty in interpreting percentage pollen diagrams from high elevations. Meaningful interpretation of the diagram is only possible in the light of parallel lines of evidence and local surface pollen data. Without such information zone 3 of the Blowdown percentage pollen diagram would be interpreted as a representative of a return to higher treelines, which is the reverse of the likely true situation.

Chapter VII Chronology

Analysis of sediments from Blowdown Lake has revealed a relatively clear and consistent pattern of environmental change within the basin during the Neoglacial. In order to relate this to the record of Neoglacial glacier advance it is necessary to establish a chronology for the Blowdown Lake core (BD4A) This chapter develops a chronology and offers initial comparisons between the results from Blowdown Lake and the Neoglacial record of glacier fluctuations.

7.1 Radiocarbon Dates

The longest (110 cm) core from Blowdown Lake (BD4A) was used to establish a chronology. The site lies within the plume of both the Mazama tephra dated ≈ 6840 BP (Sarna-Wojcicki, *et al.* 1983) and the Bridge river tephra dated ≈ 2350 BP (Mathewes and Westgate, 1983). However neither of these ash falls is apparent in the core. Radiocarbon dating was therefore required in order to establish a chronology. The core includes a section of wood at 90 - 96 cm depth. It appears that the corer penetrated cleanly through a buried log. A large quantity of wood was therefore available to obtain a conventional radiocarbon date at this depth. The resulting date was 4140 ± 100 BP (Beta 57762). In addition, an accelerator mass spectrometry (AMS) date of 1460 ± 50 BP (TO-3876) was obtained for a twig at 28 cm depth. Needles from 67 cm yielded a date of 4000 ± 90 BP (TO 3877).

The dates from 67 and 90 cm depth present a problem of interpretation. If both are correct, a sedimentation rate between the two, an order of magnitude greater than that of the rest of the core is required. This is highly unlikely, especially as laminations and sand layers between the two horizons suggest orderly sedimentation of similar character to the rest of the core. Careful analysis of possible errors associated with emplacement of the dated material within the sediment body must therefore be undertaken. The dated material from the 90 cm level was a section of a large log. Since broken trees are common in the avalanche runout zone adjacent to the lake, this is the most likely mechanism for transport of the log to the lake. It is possible that trees which were present on the surface for a considerable time were swept into the lake by later avalanches. In this case the date on the log would be anomalously old. Washing of long dead organic material into the lake is the most likely source of error in attaching radiocarbon dates to the sediment sequence. The rocks in the basin are acidic so that old carbon errors are not a concern. The most likely cause of anomalously young dates is contamination of dated material by younger organic material, for example rootlets of more recent plants. The dated material was examined for this type of contamination and none was obvious.

The dated material from the 67 - 68 cm level was five conifer needles found in a sand layer that contained considerable organic matter and charcoal. Some of the material was dark and matted, suggestive of inwashed soil materials.

Since it is unlikely that either of the problem dates is anomalously young, it is the later date which is suspect, since if the lower date was too old the dates would invert. It is suggested that the needles dated were preserved in the humic layer of a forest soil and washed into the lake at a later period during a major erosional event.

The chronology of the core is therefore assumed to be represented by dates of 1460 BP at 28 cm depth, and 4140 BP at 90 - 96 cm depth. This translates to sedimentation rates of 0.019 cm/yr for the first 28 cm of the core, and 0.024 cm/yr from 28 to 90 cm. These are of course sedimentation rates for the core, not the lake. Taking into account error introduced by compression of the whole core during extrusion this is suggestive of relatively constant mean rates of deposition throughout the core. Sand layers in the core however suggest that this orderly deposition was interrupted by occasional periods of rapid sedimentation. Compaction of the lower sediments normally produces lower apparent sedimentation rates at the base of the core. However the amount of force exerted on the core by using a chain hoist to extrude it is considerable. Up to 20 cm of compaction occurred before the core began to extrude from the corer. Such compaction would be concentrated on the unconsolidated upper sediments producing a relatively constant amount of compaction throughout the core.

In order to interpolate between the radiocarbon dates obtained, and hence date the environmental changes of interest, it is necessary to draw some conclusions about sedimentation rates to provide a basis for the interpolation. As noted the similarity of observed sedimentation rates above and below 28 cm depth suggest relatively constant sedimentation. However there are two major problems with this conclusion. The first is that the sand layers in the core suggest episodic sedimentation, and the second is that the extent of compaction, both natural, and produced during extrusion, is unknown.

The pollen concentration data presented in Figure 6.5 provide some basis for an assessment of sedimentation rates. Changes in pollen concentration are either associated with decreased sedimentation rates, or with increased production of pollen. The pollen concentration data show a major change from higher to low concentrations at around 80 cm depth. This is either associated with decreased production or with higher sedimentation rates. Other evidence of treeline shifts at this period suggest that the change is at least partially a result of changes in pollen production. At depths less than 80 cm pollen concentrations are relatively constant. There are spikes of short term variation but the background levels appear quiescent. In particular the concentrations of lowland pollen types are relatively constant. Lowland vegetation is relatively insensitive to climatic change, so variation of production during the last 4 ka should have been minimal. Therefore the observation that pollen concentrations have been reasonably stable since the period represented by sediment at 80 cm, provides some support for the suggestion that sedimentation rates have been constant over time. The presence of sand layers in the core show that sedimentation has been episodic, but it may be assumed that average rates have been relatively constant.

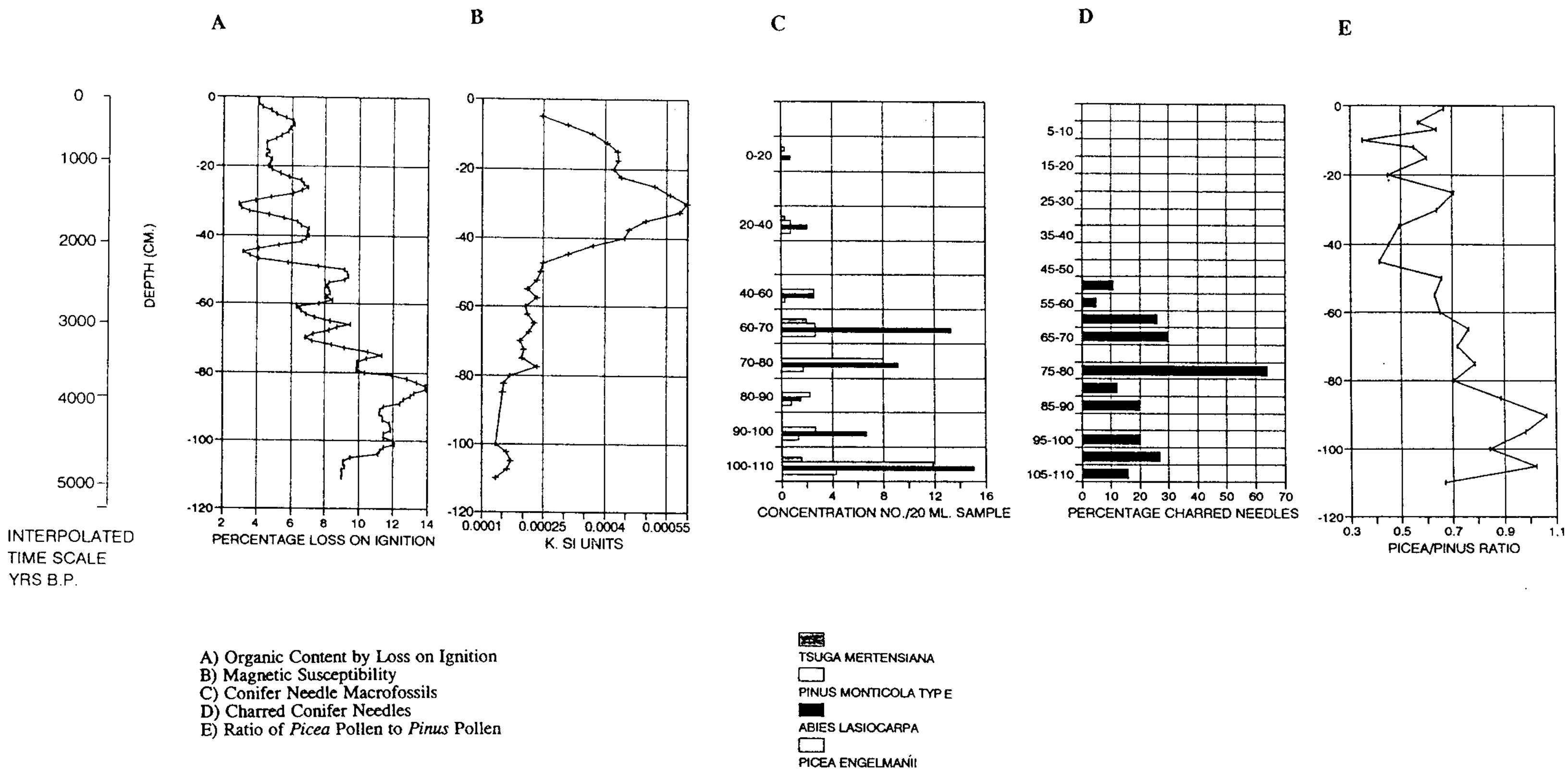
Based on the assumption of constant sedimentation rates at least in the upper 80 cm of the core a chronology for the core was developed. It was produced by linear interpolation between the acceptable radiocarbon dates. Obviously the error bars associated with dating by this technique are large, however the dates do provide a basis for discussion of the Blowdown Lake results in a regional context. Dates in the following analysis cited as a number of years BP are derived by interpolation.

7.2 Palaeoenvironmental Change at Blowdown Lake

Figure 7.1 summarises the evidence of palaeoenvironmental change from Blowdown Lake. Included on this figure is the interpolated chronology of the core. Parallel lines of evidence have been vital to interpretation of the various proxy data sources considered. The figure clearly illustrates the consistent picture of environmental change, presented by both palynological, and non-palynological lines of evidence. The evidence for, and dating of, major periods of environmental change are discussed below.

From 4600 BP until 3400-3800 BP pollen and macrofossil evidence suggest that treeline was higher than at present. Dense forest probably surrounded the lake. LOI is relatively high during this period. This probably reflects the supply of organic material to the lake from the forested catchment. Low magnetic susceptibility values may result from the organic nature of the sediment, however, they may also reflect lower input of mineral material to the lake. This implies less catchment erosion, and is consistent with stable vegetated slopes. Charred needles suggest a major fire episode at around 3600 BP It is possible that this fire hastened treeline retreat by removing live trees from the

Figure 7.1 Summary of Evidence From Blowdown Lake



zone where they may survive, but seedling establishment is unlikely. The evidence of fire slightly postdates the first indications of treeline retreat in the pollen record. The fire may therefore have been burning dead wood above treeline. Most likely a combination of effects occurred. Fire in dead wood may have spread to live trees below treeline hastening treeline retreat.

From 3400 BP until around 2500 BP pollen evidence suggests lower treeline. Gradual increase in magnetic susceptibility, and declining LOI imply a transitional period, characterised by either lower organic inputs to the lake, higher mineral input, or both. This suggests increasing erosion in the catchment.

The period from 2500 BP to the present has been one of fluctuating lower treeline. Pollen ratio data identifies two periods of low treeline. The other proxies do not detect this minor fluctuation. However minima of macrofossil concentrations, minerogenic sediment, high magnetic susceptibility values, and the cessation of deposition of charred needles, all point to lower treeline, and associated increases in erosion.

It is important to recognise the danger of circularity of arguments in palaeoenvironmental reconstruction. None of the proxy data sources is an error free measure of an objective reality. Considerable uncertainty is inevitable when attempting to make detailed reconstructions of past environments. However, the consistency of the timing of change across a range of proxy measures in this study, provides strong support for the the interpretations offered above

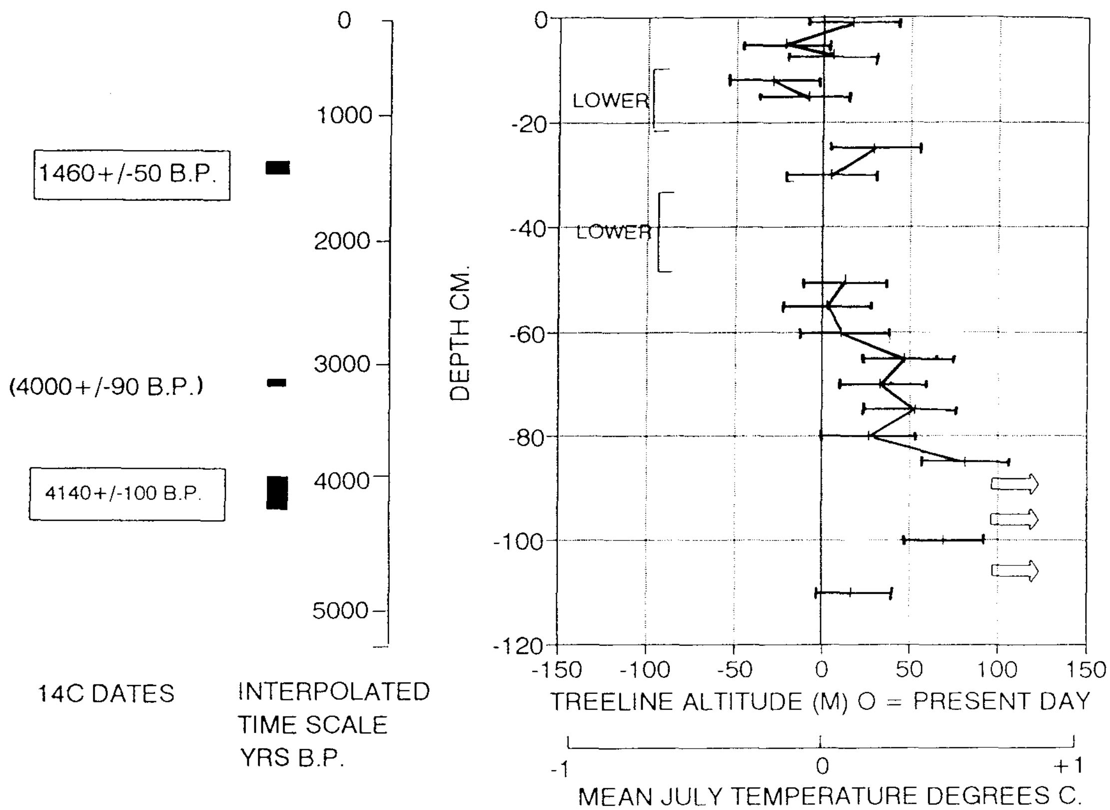
Having established a chronology for the evidence of environmental change at the Blowdown Lake site it is possible to evaluate the results in terms of the objectives of the thesis, i.e. to assess the extent to which the pollen record registers climate changes associated with periods of Neoglacial advance identified by Ryder and Thomson (1986).

Figure 7.2 shows the changes in treeline altitude derived from pollen ratios (shown in

Figure 6.5) with the addition of the interpolated timescale. Treeline is high until 3800 BP and then declines steadily. The two periods of lower treeline than present date at 2400-1500 and 1200-200 BP. These periods correlate reasonably well with the mid and late Neoglacial advances identified by Ryder and Thomson, (Tiedemann Phase, 3300-1900 BP and the Little Ice Age, 900-100 BP). The maximum of the earlier Tiedemann glacial phase was at 2500 BP The time difference between the Tiedemann phase and the earlier period of low treeline may be a function of the lag in treeline response to climatic change identified in Chapter II. However this phenomena cannot explain the difference between the dates of the later period of low treeline and Ryder and Thomsons' Little Ice Age dates. The slight non-synchroneity may simply be a result of the inexact nature of dating by interpolation. The apparent correspondence between periods of lower treeline and Neoglacial phases supports the hypothesis that with careful site selection relatively minor climatic changes may be identified in the pollen record.

The implications of the Blowdown Lake evidence for understanding of regional Neoglacial climates are discussed in Chapter IX.

Figure 7.2 Palaeo-treeline Altitudes at Blowdown Lake



Error bars = one standard error.

Chapter VIII Little Ice Age Equilibrium Line Depression in the Southern Coast Mountains of British Columbia

8.1 Introduction

The evidence of treeline variation presented in the previous chapter clearly identifies a transition from warmer to cooler climates around 4000 BP. Chapter II hypothesised that calibrated pollen records may allow quantification of Neoglacial climatic deteriorations. The pollen evidence presented in Chapter VI supports this assertion but lack of suitable sites for surface pollen sampling prevented calibration of cooler periods.

The difficulty of obtaining quantitative estimates of palaeoclimate from records of glacial fluctuation has already been noted (section 1.3). However such a procedure holds the prospect of deriving estimates of climate changes during the cooler periods for which calibration of the pollen record was not possible. Therefore, this chapter attempts to derive at least a first order approximation of the extent of the cooling/wetting of climate directly from the geomorphic evidence of ice advance. The destruction of evidence of early and mid-Neoglacial phases by the extensive Little Ice Age advance, limits such an analysis to the latter advance. However, since Denton and Karlen (1973) suggest a common forcing mechanism for major periods of Holocene glacial advance it is reasonable to assume, at least at the outset, that all the Neoglacial periods are analogous in terms of climatic changes.

8.2 Glaciers and Climate

Glacier mass balance is a function of accumulation and ablation. The primary controls on these phenomena are precipitation and temperature. Locally, factors such as drifted or avalanched snow may enhance accumulation, whilst local thermal regimes influenced by topography may not be regionally representative. Nevertheless, alpine glacier fluctuations are primarily a response to changes in temperature or precipitation. For example, Porter (1977) showed that in the Cascade Mountains of Washington state, 90% of the variance in glaciation threshold[1] is explicable in terms of accumulation season precipitation and mean annual temperature. Reynaud (1980) demonstrated that summer temperatures were the major control on mass balance in the European Alps. Grove (1988) notes that data from Scandinavia suggest precipitation is the dominant component of the mass balance in maritime regions, but that its importance diminishes with increasing continentality.

Bradley (1985) notes that a wide range of climatic conditions may produce the same mass balance, and hence the same dynamic glacier response. Therefore interpretation of glacial advance or retreat in terms of specific climatic parameters is difficult. Separating the effects of temperature and precipitation is particularly problematic. If sufficient modern climate data are available, an envelope of possible temperature and precipitation conditions for a given glacial advance or retreat may be calculated (Porter, 1977), however such an approach is not useful in a region with sparse modern climate data.

Increases in *Tsuga* pollen deposited at Blowdown Lake in the last 1500 years suggest increasingly moist climate during the Late Neoglacial (see 6.2.1.2). However, Mathews (1951) argues that Little Ice Age glacier fluctuations in the Garibaldi area are primarily associated with temperature change because he found no significant long term change in precipitation in climate records stretching back to 1859. Mathews suggestion is that secular precipitation variation has been a secondary influence on glacier retreat (since the climax of the LIA). In contrast Deslosges (1987) demonstrated that at three monitored glaciers in the southern Coast Mountains (Bridge, Place, Sentinel) changes in mass balance in the period of instrument record are most strongly associated with changes in winter precipitation. He argues that the maximum LIA advance in the

1 Defined as the mean of the altitude of the lowest glaciated peak and the highest unglaciated peak within a region. An approximation of the regional level at which ice is maintained under climatic control.

Northern Pacific ranges of the Coast Mountains was due to cooler *and* wetter climatic conditions.

Dendrochronological evidence of summer precipitation since 1600 is available for sites 80 km to the east of Blowdown Lake (Jones, 1988). These records suggest that precipitation was average or above average for most of the 18th century and until 1825. A drier period lasted from 1825 to 1870, and was succeeded by wet conditions from 1870 until 1920. The complete records show alternating periods of higher and lower precipitation. No long term monotonic trend is apparent. Figure 8.1 compares tree ring records of LIA precipitation (Jones, 1988; Deslosges, 1987) with the ages of dated LIA terminal moraines in the Coast Mountains (Ryder *et al.,* 1981; Deslosges, 1987). Whilst this figure does not conclusively demonstrate a link between wetter conditions and glacier advance, a reasonable correlation is apparent.

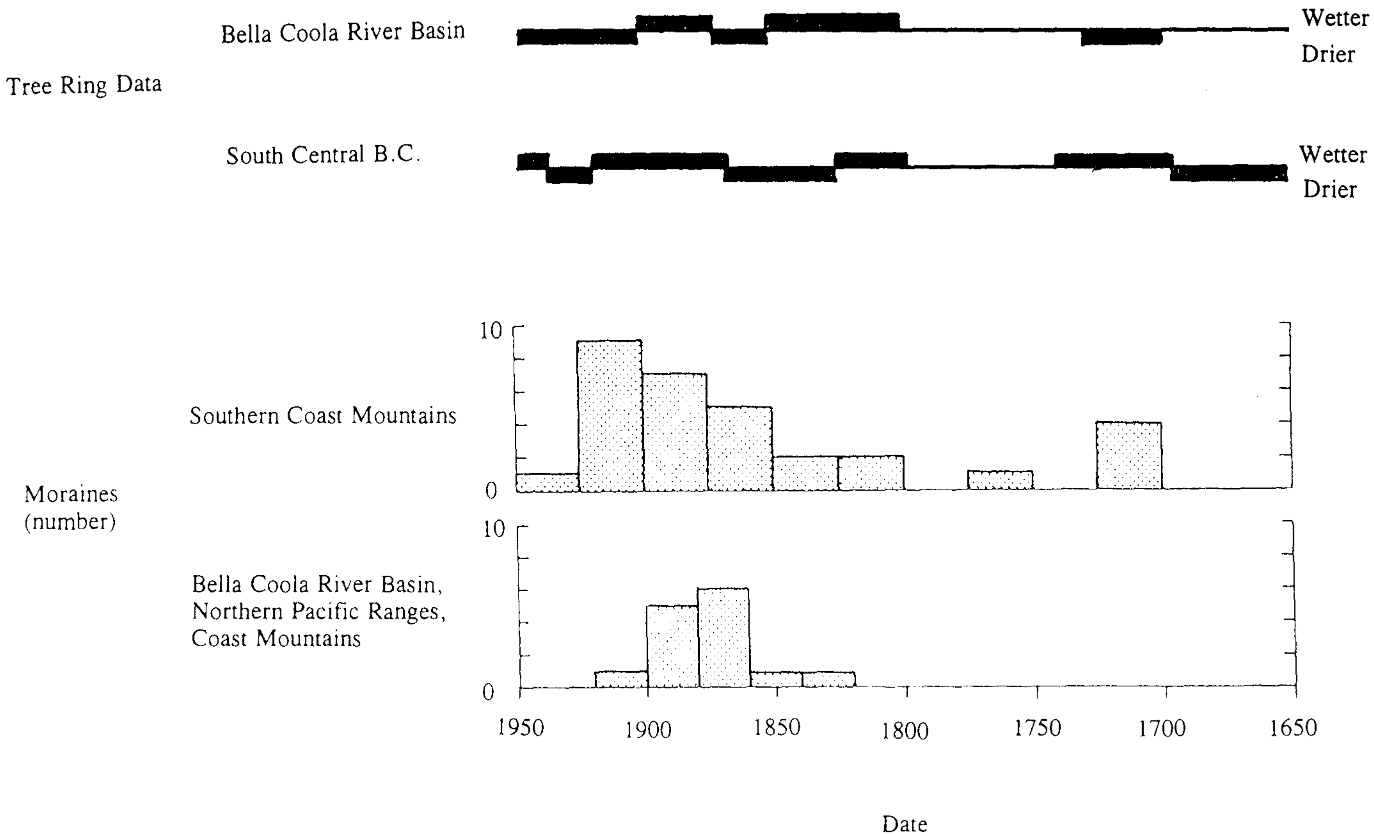
The nature of climatic change during the LIA is uncertain, however it is argued that the balance of the evidence suggests that LIA conditions were as wet, or wetter than the present. Making such an assumption allows interpretation of LIA glacier advances in terms of a maximum temperature depression.

Reconstructions of climate based on evidence of glacial fluctuation have frequently focussed on changes in equilibrium line altitude (ELA)[2] (e.g. Sissons, 1980; Porter *et al*, 1986; Clapperton, 1987). The equilibrium line altitude is defined a the elevation on a glacier at which accumulation equals ablation such that the net mass balance is zero. As such it is clearly a climatically determined parameter. However it is not sufficient to simply determine past and present ELA's at a site, and infer a climatic change from the difference.

Glacier mass balance also responds to changes in the very local climate at the glacier surface, as well as to factors not directly under climatic control such as avalanching and blowing snow. Further, glaciers exhibit variable lag times between a

2 The ELA closely corresponds to, but is distinct from the firn line (Porter, 1975). Similarly glaciation limit is a related theoretical surface which normally parallels the equilibrium line, but at a higher altitude (Porter, 1977).

Figure 8.1 Comparison of Tree Ring "Wetness" Records With Dates of Maximum LIA Advance



Moraine plot after Deslosges (1987) Data from Deslosges (1987) and Ryder *et al.* (1981).

Bella Coola tree ring data after Deslosges (1987), Southern B.C. tree ring data after Jones (1988).

change in mass balance and advance/retreat of the snout (Porter, 1981). Evidence also suggest that some surging glaciers may advance and retreat apparently independently of climate (Mann,1986). Over emphasis of local variation can be prevented by integration of regional patterns of ELA change. Where strong climatic gradients exist, such integration may be achieved by plotting transects of ELA across an area parallel to climatic trends. The difference between the lines representing past and present ELA surfaces may then be interpreted in climatic terms. In addition, changes in the slope of the surfaces may reveal past changes in the steepness of the climatic gradient. Modern glaciation limits have been mapped for southern B.C. (Ostrem, 1966). Ostrem (1973) also briefly considered modern ELA's in the Coast Mountains. Ostrems work on glaciation limit was refined and expanded for a small area of the southern Coast Mountains by Evans (1990). However, the author is unaware of previous attempts to reconstruct palaeo-ELA's in the region.

8.3 Methods and Preliminary Results

All methods of estimating ELA's have both advantages and disadvantages. The particular methodology utilised in this study evolved as the nature of the problem in the study area was illuminated. The methodology is discussed below, along with preliminary results which led to modifications to the approach taken.

8.3.1 Estimating Equilibrium Line Altitudes

Exact calculation of ELA necessitates detailed glaciological measurement to determine the glacial mass balance. Modern ELA's correspond closely with snowline elevation at the end of the ablation season (Meier and Post, 1962). As such they may be easily estimated in the field, or from air photographs taken in the appropriate season (e.g. Ostrem, 1973). However, estimates of ELA based on the transitory snowline, on a single air photograph, are specific to a particular year's meteorological conditions. They may not be fully representative of a climatically controlled mean. Use of several years photographs to obtain mean values is preferable. In addition, utilising air photographs is obviously not an option for assessing LIA ELA's. Therefore a variety of indirect techniques for estimation of the ELA have been developed. These include the glaciation limit (Ostrem, 1966), median altitude of small glaciers (e.g. Porter, 1975), various ratios of glacier toe to headwall altitudes (THAR) (Meierding, 1982), altitude of lateral moraines (e.g. Andrews, 1975), cirque floor altitudes (e.g. Andrews, 1975), and assumed ratios between the accumulation area and the ablation area of a glacier (accumulation area ratio technique, AAR) (Meier and Post, 1962).

A detailed, critical review of these techniques, and assessment of relative accuracies is provided by Meierding (1982) who demonstrated that in the Colorado Front Ranges, the smallest departures from regional ELA trend surfaces were obtained using AAR's of .65 or THAR of .4. A recent study from Jostedalsbreen (Torsnes *et al.* 1993) suggests that the AAR technique gives superior results where glaciers do not have a regular area/altitude distribution. The AAR technique has also been modified to incorporate variable hypsometry (Leonard, 1984).

8.3.1.1 Modern ELA's in the Southern Coast Mountains

In order to estimate LIA depression of equilibrium lines, it is first necessary to describe the pattern of the equilibrium lines of modern glaciers. Meierding (1982) notes that the AAR technique of estimation is theoretically superior to THAR since it integrates the surface area, as well as elevation and ice gradient, of the glacier. Initially therefore for the present study, it was decided to determine the equilibrium lines of glaciers from a single map sheet based on accumulation area ratios. This method was chosen since in order to make valid comparisons with the LIA ELA's, a technique applicable to both modern and palaeoglaciers was required.

The approach taken in this study was to determine equilibrium lines for glaciers along a east-west transect, from the coast (west) to Blowdown Lake (east). Initially equilibrium lines were determined for all the glaciers on the 1:50000 NTS map of Clendinning Creek (92 J5). Measurements were made of simple glaciers and icefield glaciers. The ice divides in the latter were drawn in on the basis of the ice surface contours. The area between each 100 ft. contour on the map was determined for each glacier by digital planimetry. This gives a potential error of $\pm$ 50 ft. which is small given the range of values. Greater error is due to operator error during planimetry, such errors were usually less than 5% but occasionally up to 10%. Hypsographic curves were then constructed for each glacier. Following Leonard (1984), glaciers were classified as icefield type, normal, or piedmont type, according to hypsometry. Accumulation area ratios of 0.75, 0.65, and 0.5 respectively were used to calculate the equilibrium lines for glaciers of each of these types (Leonard, 1984).

Given the strong climatic gradient across the Coast Mountains from wet in the west to drier in the east, an eastward rise in the mean ELA would be expected. At a much smaller scale Ostrem (1966) showed that the glaciation limit rose from the southwest to the North east across the coast range. Evans (1990) produced similar findings for a smaller area which included map sheet 92/J5. The results of the calculations of ELA are presented in Figure 8.2. The calculated ELA's vary in the range 1250-2250 m The considerable variability in the data, and the lack of an obvious trend suggest that local variation is dominant over broader climatic trends in this map area. Figure 8.3 is a plot of the data from Figure 8.2 along an east-west transect[3]. Again, the strong variability of the data is clear. It is worth noting that the method adopted in this study is to fit a line through the original data. This preserves the scatter in the data which is real and of interest. Ostrems study removed much of this variation

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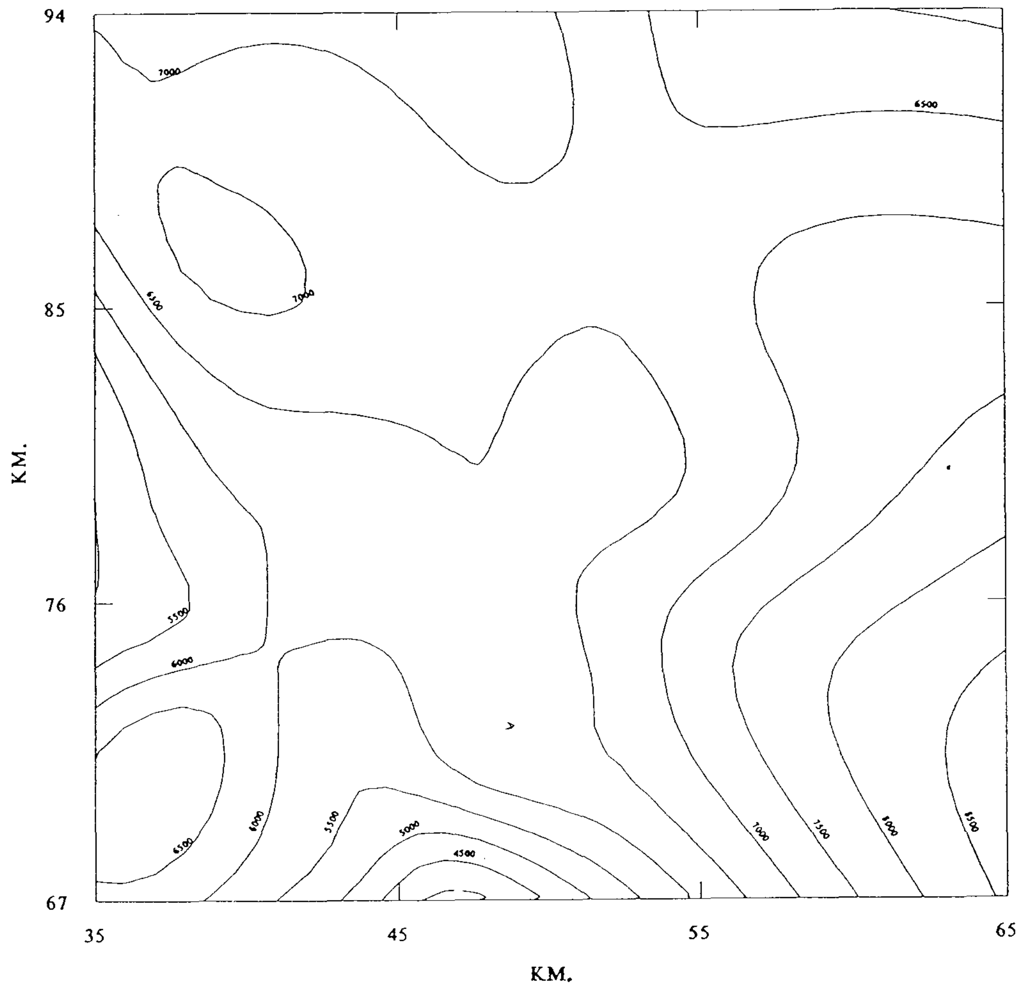
3 Points are plotted against and east-west transect not, northeast-southwest. If ELA parallels glaciation limit across the region the latter is the true regional gradient. However, this study is based on just five 1:50000 map sheets representing a transect of the Coast Mountains. The range of latitude covered by a single map sheet is only 15'. Therefore the dominant axis of variation over the five map sheets will be east-west.

Figure 8.2 Equilibrium Line Altitudes for Selected Glaciers from Map Sheet 92 J/5 - Calculated by the Accumulation Area Ratio Method



Origin 124° W. 50°15' N.

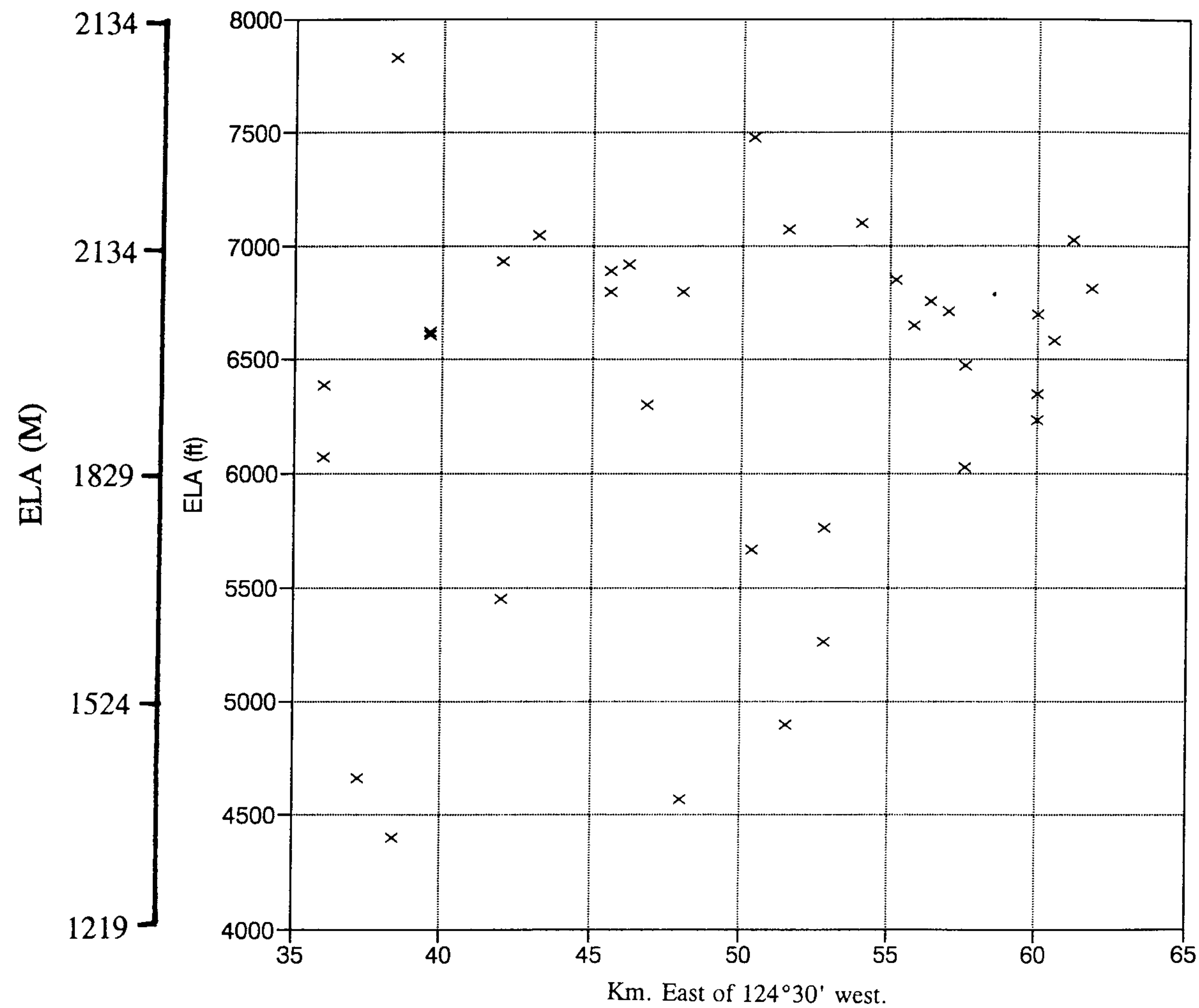
Elevations in Ft.

Metric Equivalents

Feet	Metres
4500	1219
5000	1371
5500	1524
6000	1676
6500	1829
7000	1981
7500	2134
8000	2438
8500	2591

Figures in chapter VIII are drawn with Feet as the unit of measurement for altitudes. This is done in order to allow easy comparison with NTS topographic maps. Metric equivalents are given

Figure 8.3 ELA's of All Measured Glaciers, Map Sheet 92 J/5 - Accumulation Area Ratio Method



by averaging glaciation level by mountain group, contouring the data, and then plotting a transect from the contoured data. In this study in order to attempt to isolate the source of local variation, various subsets of the data were plotted. The data was divided according to aspect, size of glacier, and hypsometry. Figures 8.4, 8.5, and 8.6. The fact that each subdivision of the data retains much of the variability of the parent data set suggests that a further influence may be controlling the local variability. One possibility is that for some glaciers avalanched snow represents a significant portion of winter accumulation. Therefore glaciers separated by rock steps may form a single glacial system. In order to test this hypothesis, the topographic location of the modern glaciers was carefully assessed and any sites at which substantial avalanching might occur were excluded. Primarily, these sites were glaciers potentially fed by ice falls from other glaciers above, and valley glaciers in deep cirques, surrounded by long avalanche slopes. Removing glaciers according to these criteria involved elimination of several anomalously

low glaciers. Once these potentially avalanche fed glaciers were excluded from the analysis, variability in ELA was reduced but not eliminated (Figure 8.7). This tends to confirm the hypothesis that the eliminated glaciers were avalanche fed.

Two important considerations were illuminated by the preliminary analysis of map sheet 92 J/5.

1) Planimetry from topographic maps and construction of hypsographic curves for every glacier is an extremely time consuming process. Therefore, the possibility that the THAR method might adequately represent regional ELA's was investigated. The THAR method works best for small, simple glaciers of normal hypsometry (Meierding, 1982; Torsnes, 1993) Therefore, toe to headwall altitude ratios of 0.35, 0.4, 0.45, and the median altitude (THAR 0.5) were calculated for the subset of glaciers on the mapsheet meeting these conditions. It is realised that this subset of glaciers in the region will not necessarily truly represent the theoretical regional ELA gradient.

Figure 8.4 ELA's for Map Sheet 92 J/5 (AAR Method) - Varying Aspect

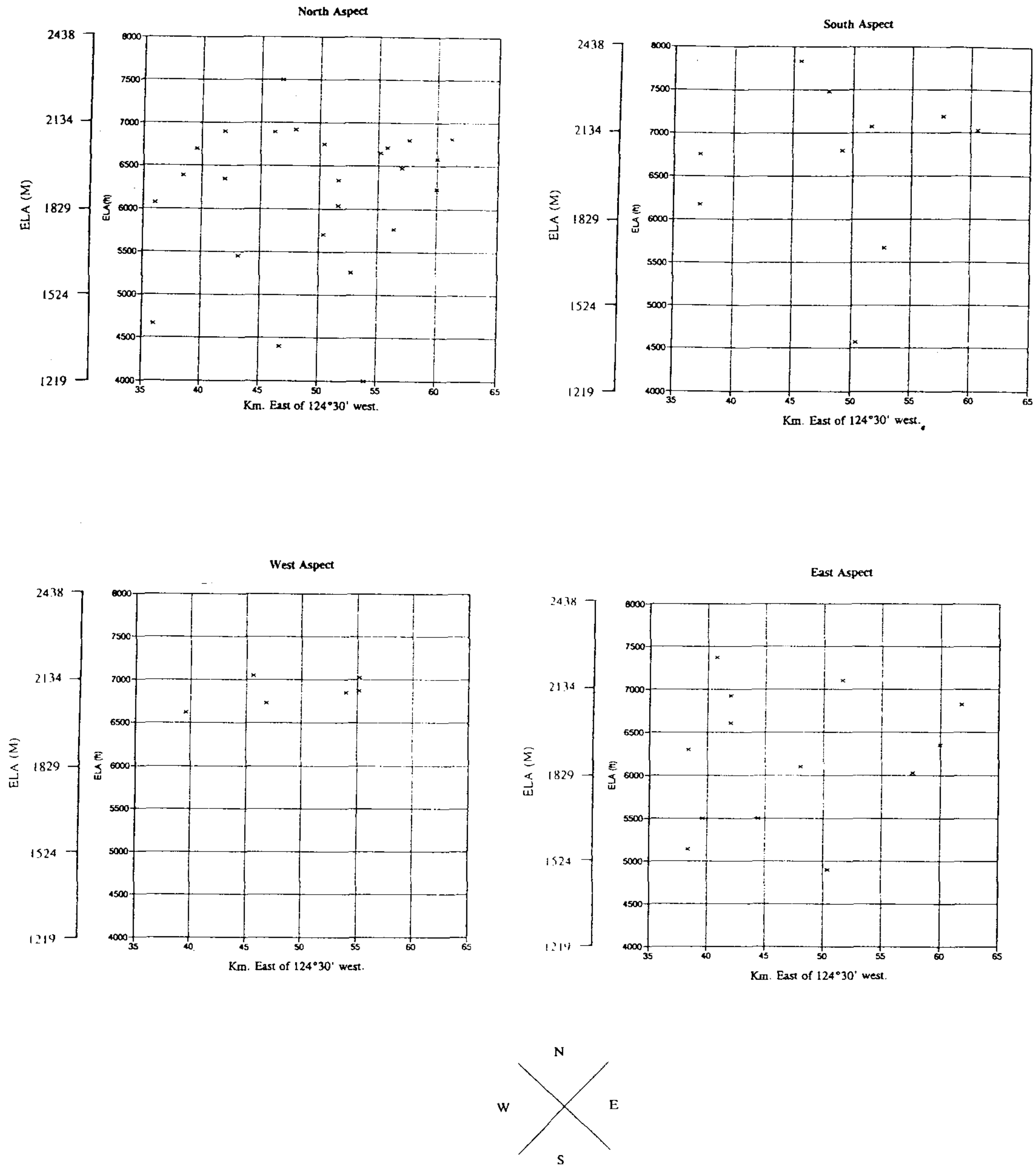


Figure 8.5 ELA's for Map Sheet 92 J/5 (AAR Method) - Varying Hypsometry

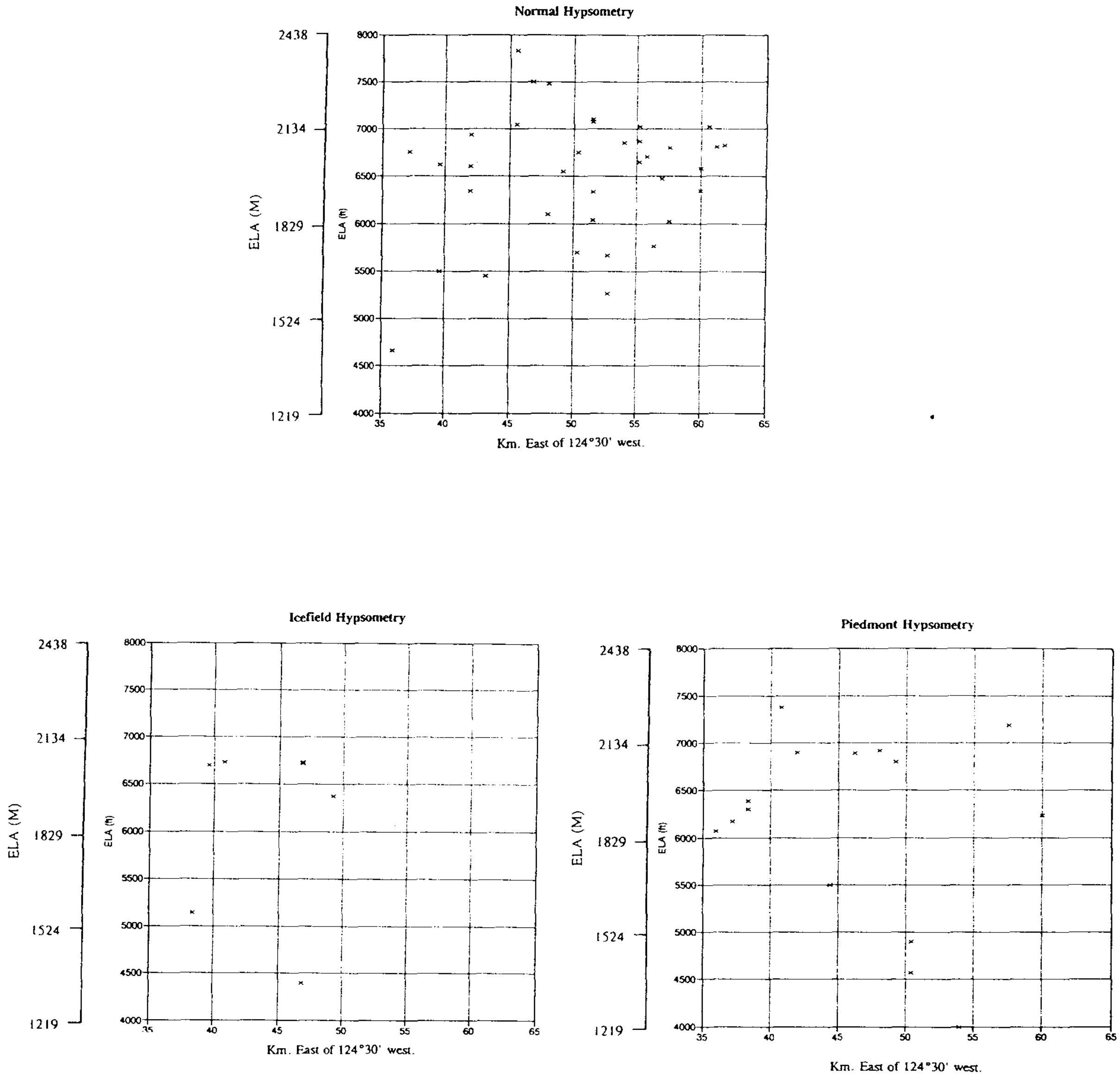
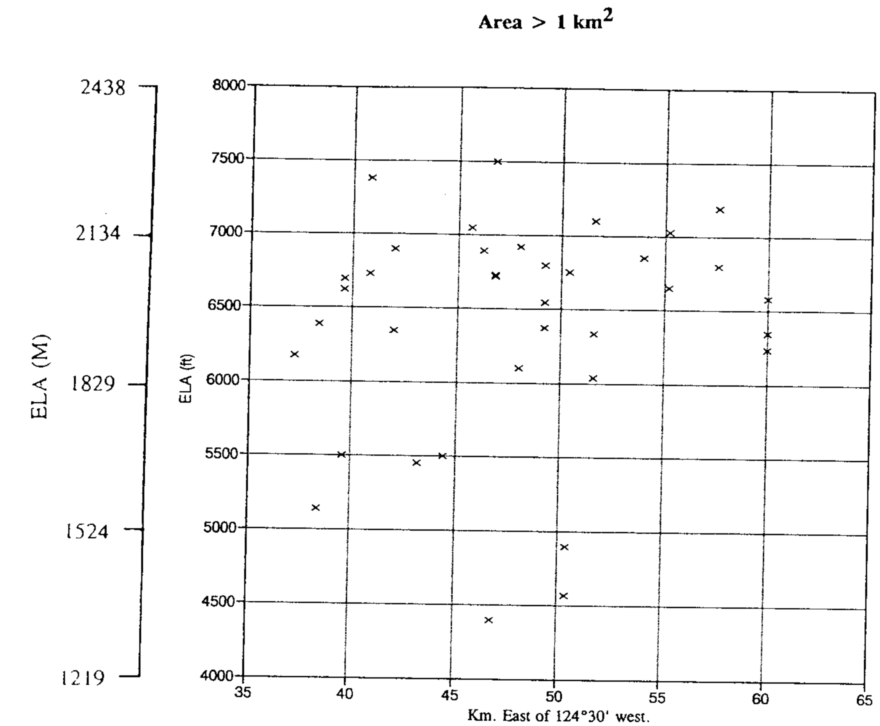


Figure 8.6 ELA's for Map Sheet 92 J/5 (AAR Method) - Varying Size



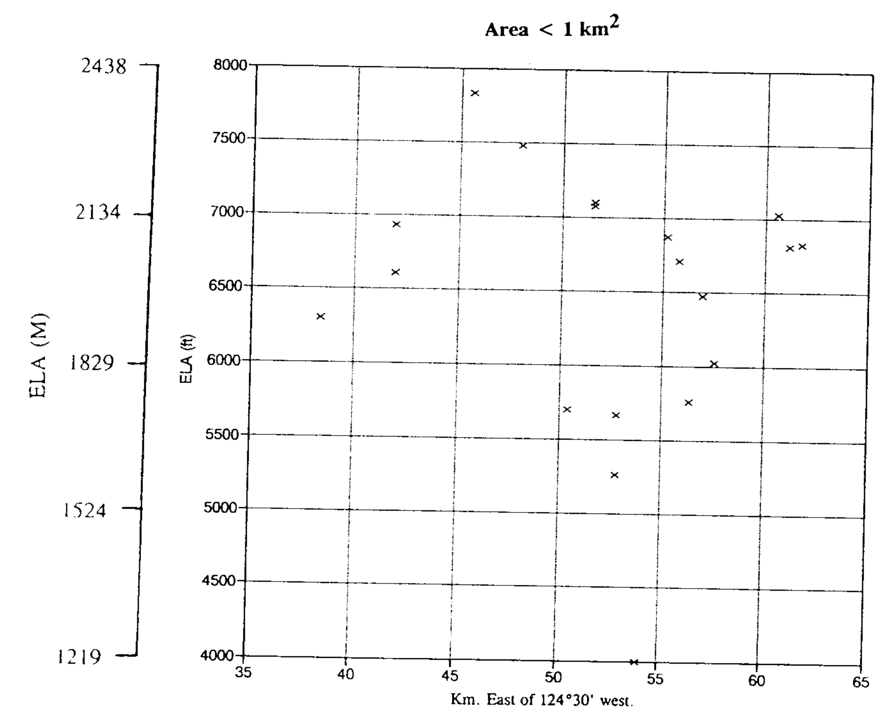
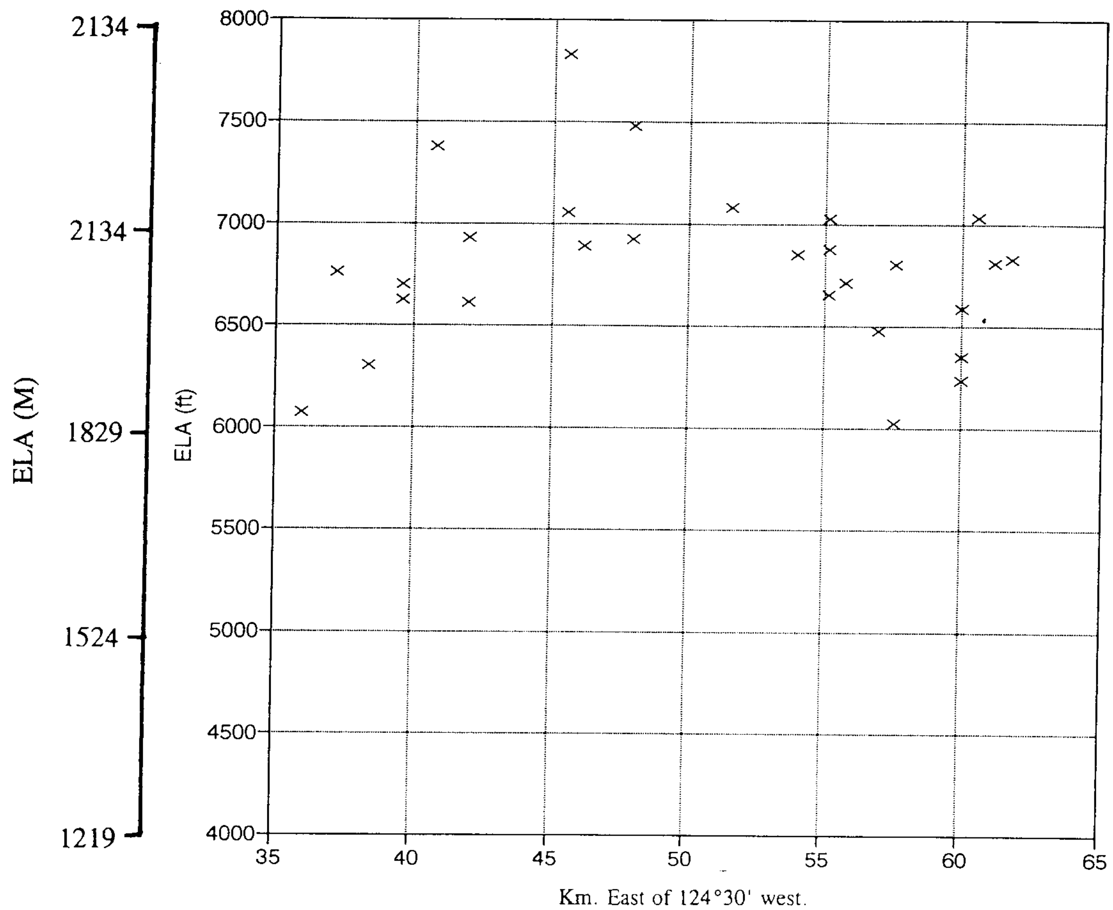


Figure 8.7 ELA's of Glaciers Assumed to have No Significant Avalanched Component of Mass Balance. Map Sheet J5 AAR Method



However the principal phenomenon of interest in this study is the extent of depression of ELA's during the Little Ice Age. Therefore, as long the glaciers for which LIA ELA's are estimated, are a representative subset of the glaciers for which modern ELA's are estimated, the results will be relatively correct. The various THAR's were correlated with the ELA's derived for the same glaciers from the accumulation area ratio technique with the following results:

THAR 0.35 vs AAR.	r = 0.869
THAR 0.40 vs AAR.	r = 0.874
THAR 0.45 vs AAR.	r = 0.874
THAR 0.50 vs AAR.	r = 0.867

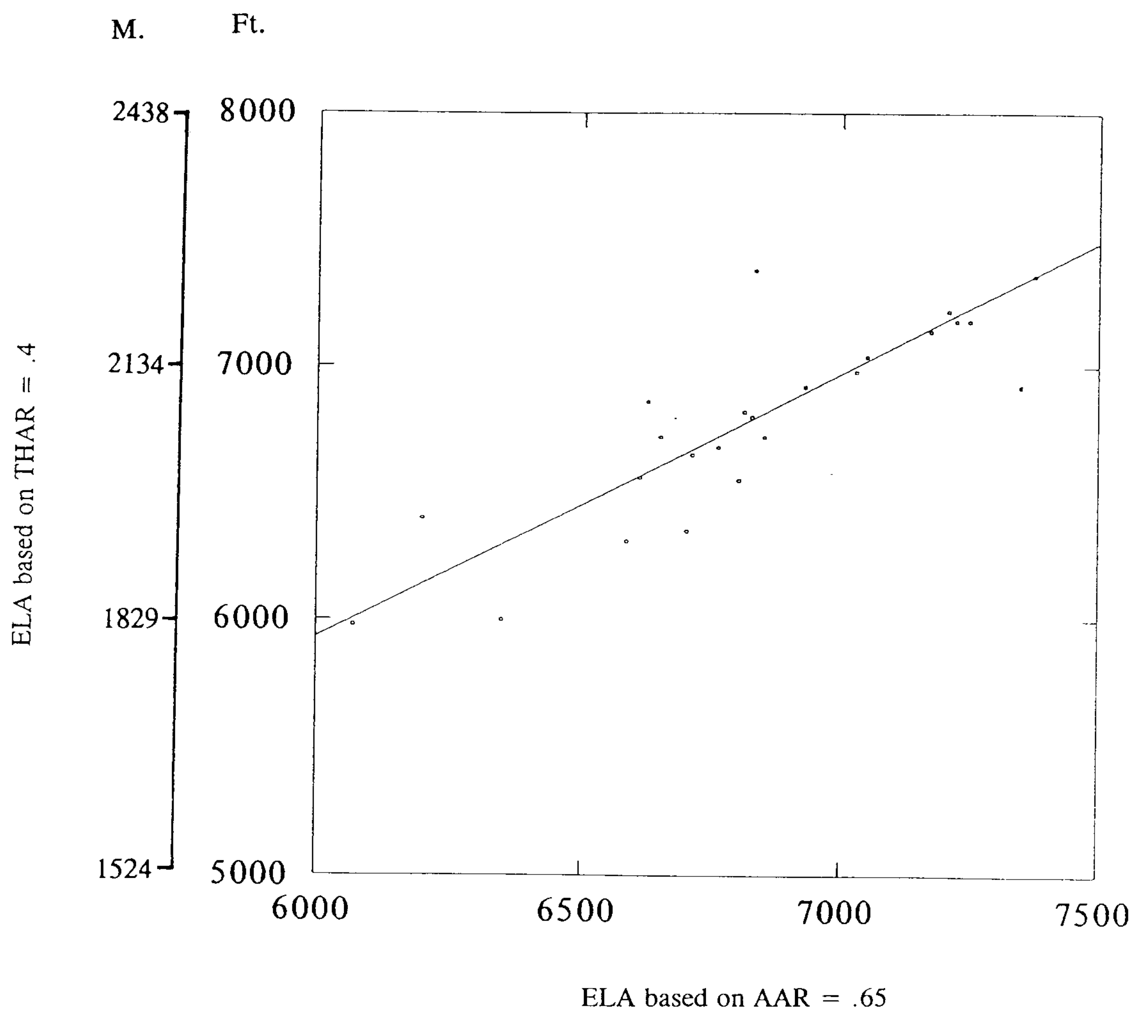
Clearly, a range of THAR values provide a good approximation to ELA values derived by the AAR method. Because of the relatively low altitudinal range of many of the glaciers in the region the absolute difference between the various THAR values was small. It was therefore determined, that for the remaining map sheets THAR = 0.4 would be used to estimate ELA. A ratio of 0.4 was selected following Meirding (1982) who used THAR=0.4 in a comparison of methods for estimation of ELA, in the Colorado Front Range[4]. The relation between AAR = 0.65 and THAR = 0.4 for glaciers on the J5 map sheet is illustrated in Figure 8.8.

2) The second point clearly demonstrated from the initial analysis, is that ELA in the study area is highly variable. At the large spatial scales than the broad region examined by Ostrem (1966), local factors strongly influence the altitude of the equilibrium line. The variability of ELA, even for glaciers carefully selected to avoid local avalanche influence, is significant. The range of this variability is of similar magnitude to the depression which might be expected during the LIA[5]. Only a proportion of the modern glaciers have well preserved moraines or trim lines permitting reconstruction of LIA

4 It is worth noting however, that close inspection of Meirdings results for a similar procedure, comparing various THAR's, also shows that the estimated ELA's are relatively insensitive to the ratio selected.

5 Based on the estimated 1°C. LIA temperature depression for the Olympic Peninsula (Porter, 1981; Burbank, 1981; cited in Grove, 1988), and on a local lapse rate of .7°C./100m.(see appendix 1)

Figure 8.8 Relation Between ELA Determined by AAR = .65 and THAR = .4, Map Sheet 92 J/5

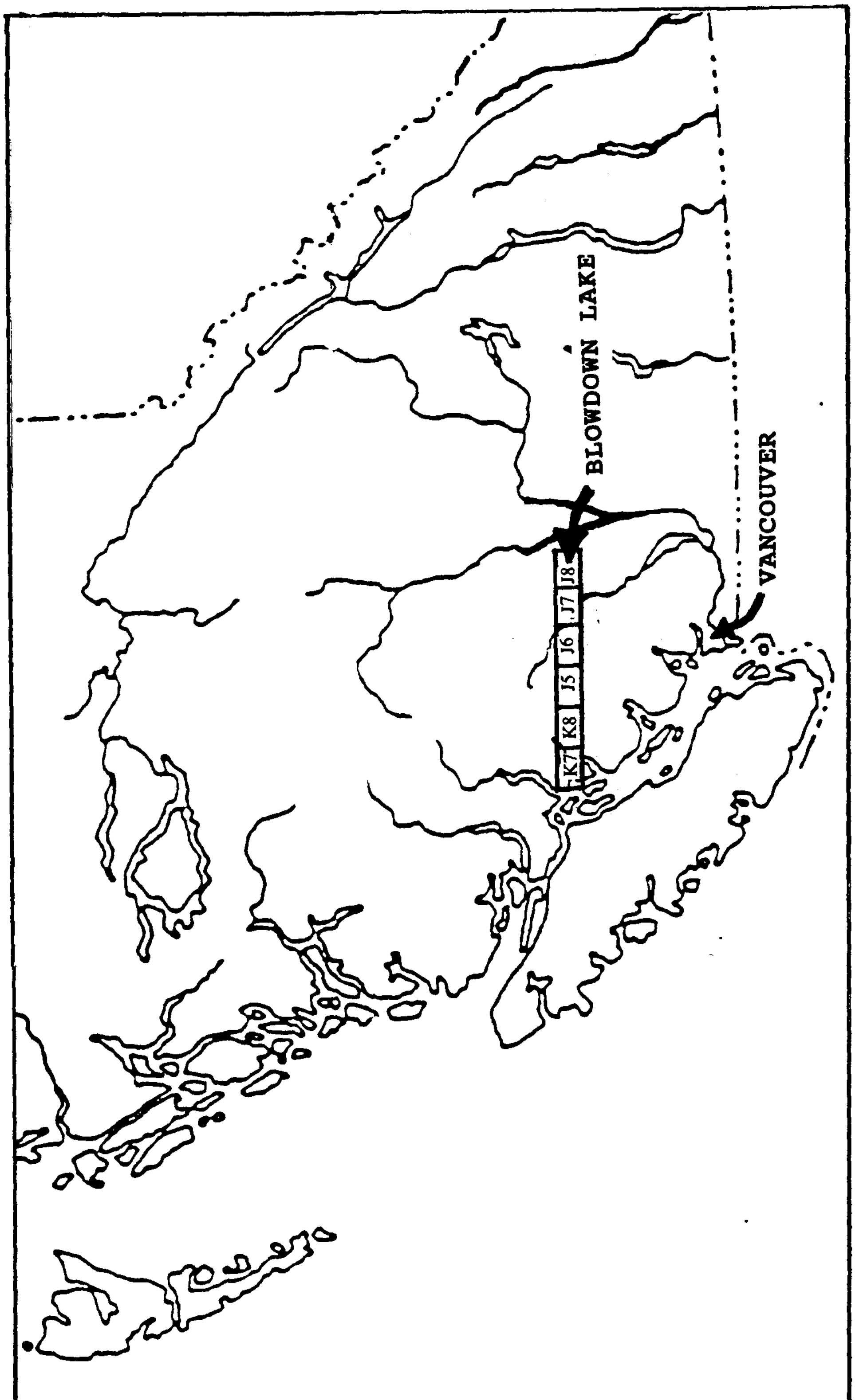


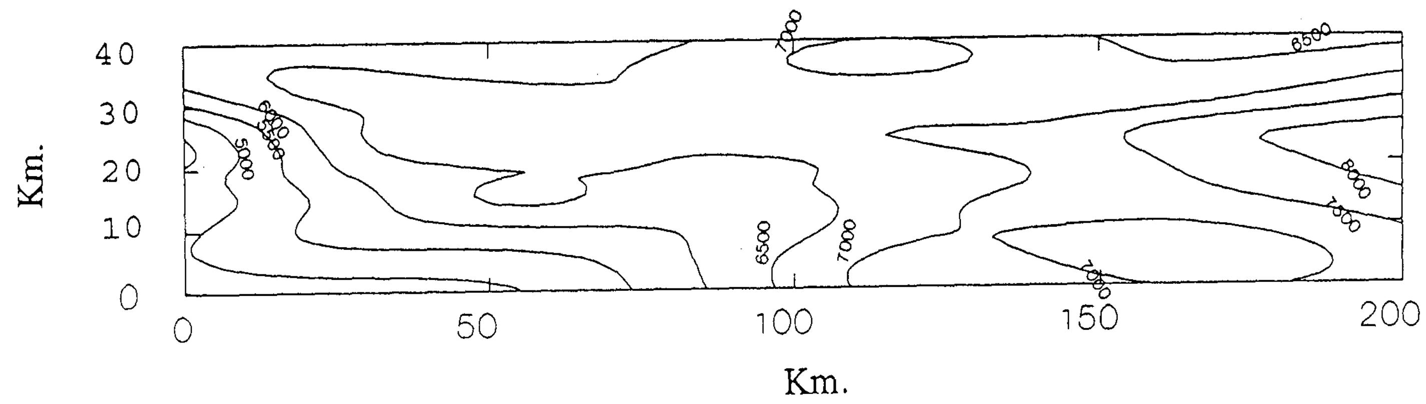
ELA. Therefore, because of the scatter exhibited by modern ELA's, comparison of the LIA ELA gradient associated with these glaciers, to the modern gradient representing all glaciers, would be misleading. It was resolved that the past and present ELA surfaces used in the final comparison should only be derived from glaciers for which the LIA extent had clearly been identified.

The map sheets selected for study were the Duffey lake 1:50000 sheet (92 J8) and the four map sheets to the west as far as the coast (J7, J6, J5, K8; Figure 8.9). In fact, the coastal map sheet is K7, but there is no permanent ice marked on this sheet. Spot checks were made between air photographs and the maps to assess the quality of the photogrammetry. At this stage some "phantom glaciers" were excluded from the data set. In general the glaciers were not well mapped, in particular the south west corner of map sheet 92 J/5 includes several errors. It is worth reiterating Evans (1990) point that checking the photogrammetry with air photographs is important for map based glacier studies in the region. The THAR technique of ELA estimation requires that the glaciers considered be simple in form, and of normal hypsometry. Unfortunateley, without constructing hypsographic curves, the hypsometry cannot be estimated quantitatively. As noted above this is excessively time consuming. Therefore on each map all the small, simple glaciers with no apparent influence of avalanching, were selected. Glaciers which obviously had icefield or piedmont hypsometry were excluded. It is important to note the assumption here that smaller glaciers are more likely to have normal hypsometry. This is simply due to the fact that glaciers with lesser extent are less likely to flow over major topographic discontinuities. The equilibrium line for each glacier was then calculated using a THAR of 0.4.

Figure 8.10 illustrates the pattern of ELA's derived by this method. The contour plot was derived by distance weighted least squares interpolation between the points. It is clear that there is a general rise in ELA from west to east. In the western half of the plot, the slope of the ELA surface is close to the southwest-northeast slope identified by Ostrem (1966). Figure 8.11 shows the same plot based only on those glaciers with LIA

Figure 8.9 Location of Studied Map Sheets





Elevations in Ft.

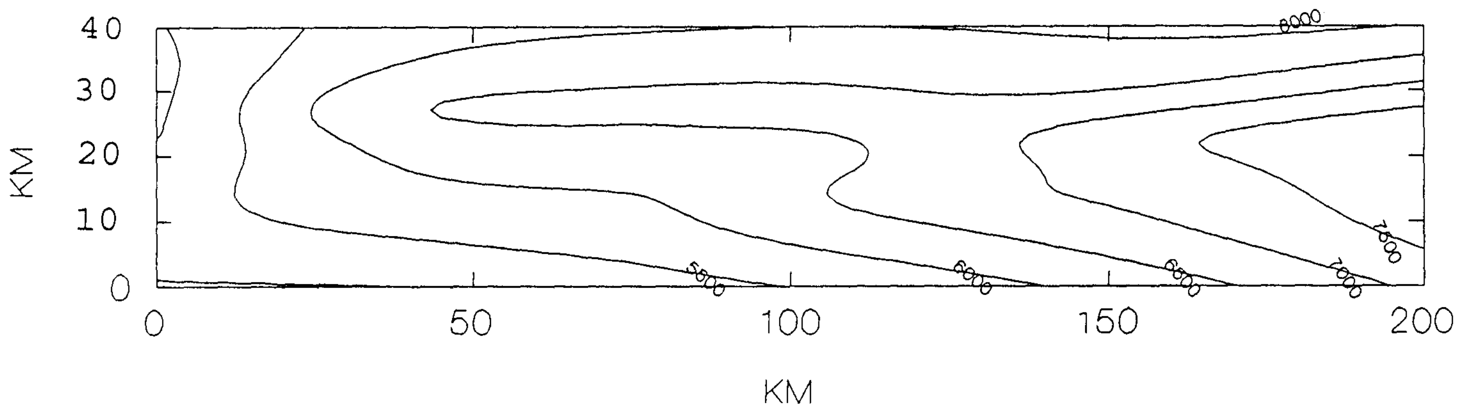
Origin 124°30' West, 50°15' North.

Metric Equivalents

Feet	Metres
4500	1219
5000	1371
5500	1524
6000	1676
6500	1829
7000	1981
7500	2134
8000	2438
8500	2591

Figure 8.10 Contour Plot of Modern ELA's Estimated by the THAR Method

Figure 8.11 Contour Plot of ELA's For Glaciers With Evidence of LIA Extent, Derived by THAR Method



Elevations in Ft.

Origin 124°30' West, 50°15' North.

Metric Equivalents

Feet	Metres
4500	1219
5000	1371
5500	1524
6000	1676
6500	1829
7000	1981
7500	2134
8000	2438
8500	2591

moraines. The west-east slope is clearer here. However the decreased local variation is most likely due to the lower number of data points and hence greater smoothing by the contouring routine. The ridge like form of the contours is also a function of the low number of points and the contouring routine. Figures 8.12 and 8.13 are plots of ELA against distance east of 124°30' West. Figure 8.12 plots all selected glaciers, Figure 8.13 those with LIA moraines. The best fit line shows the east-west gradient of ELA for small, simple, modern glaciers of apparently normal hypsometry. The best fit line on Figure 8.13 has a slightly steeper slope than that on Figure 8.12. However given the scatter of the data, no conclusion can be drawn as to the representativeness of the subset of glaciers with preserved moraines. As noted previously it is relative change in the ELA surface elevation is of interest rather than absolute values. The results should therefore be meaningful as long as comparable datasets are used for both LIA and modern ELA's. The data from 6.13 were broken down by aspect and replotted, in an attempt to account for some of the remaining scatter (Figure 8.14). All of the glaciers with clear evidence of LIA advances preserved had a northern or easterly aspect. The two groups preserve the scatter around the best fit line of the original data set. However the predominance of these two aspects is of interest. Preservation of moraines, or trimlines at sites of northern and eastern aspect is intuitively reasonable, since shading at the warmest part of the day will lower local temperatures. The predominant trajectory of storm tracks across the coast mountains is southwest to northeast. The glaciers with northerly or easterly aspects are therefore on the lee slopes. The combination of shading, and lee slope accumulation of snow produces probably produced strongly positive mass balances in glaciers of north and east aspect during the LIA. These glaciers would therefore have advanced vigourously producing clear moraines.

Figure 8.12 ELA's of Modern Coast Mountain Glaciers - THAR Method

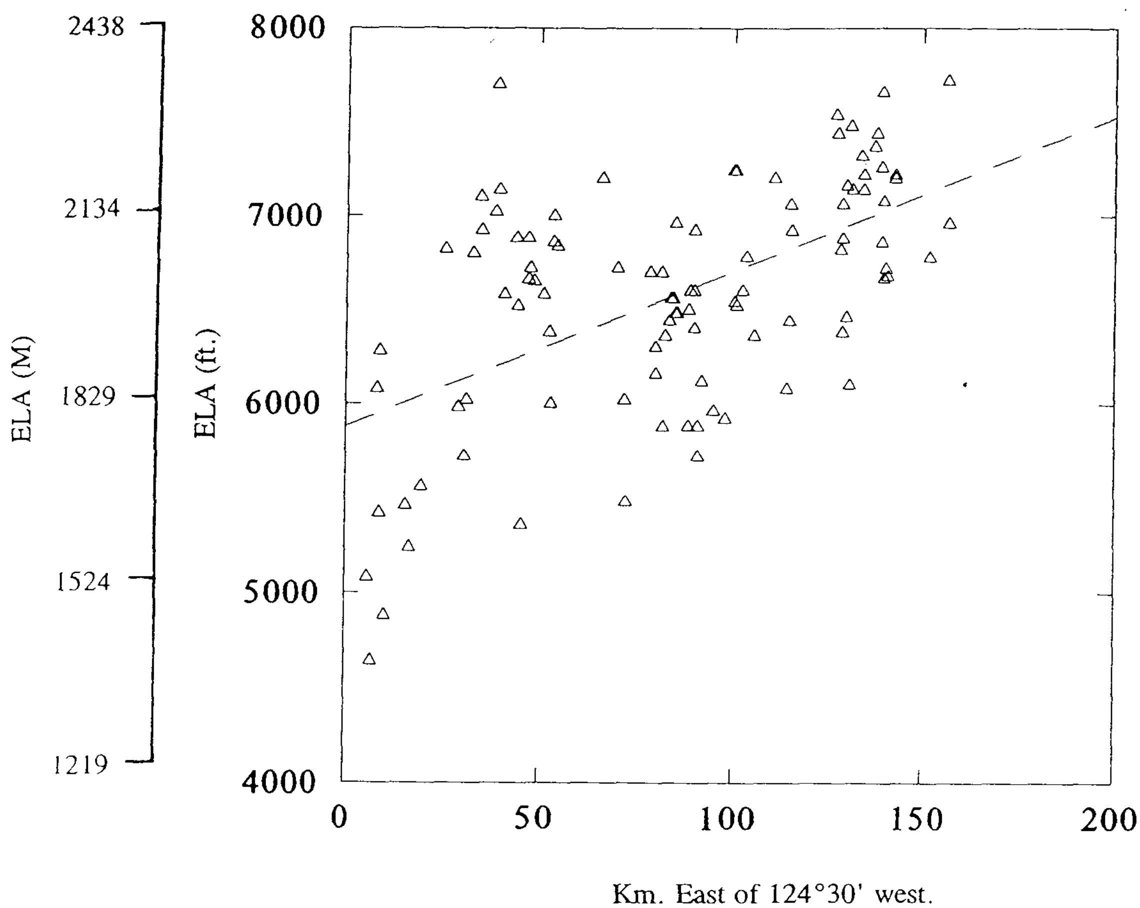


Figure 8.13 Ela's of Modern Coast Mountain Glaciers for Which the Extent of the LIA Advance Was Determined

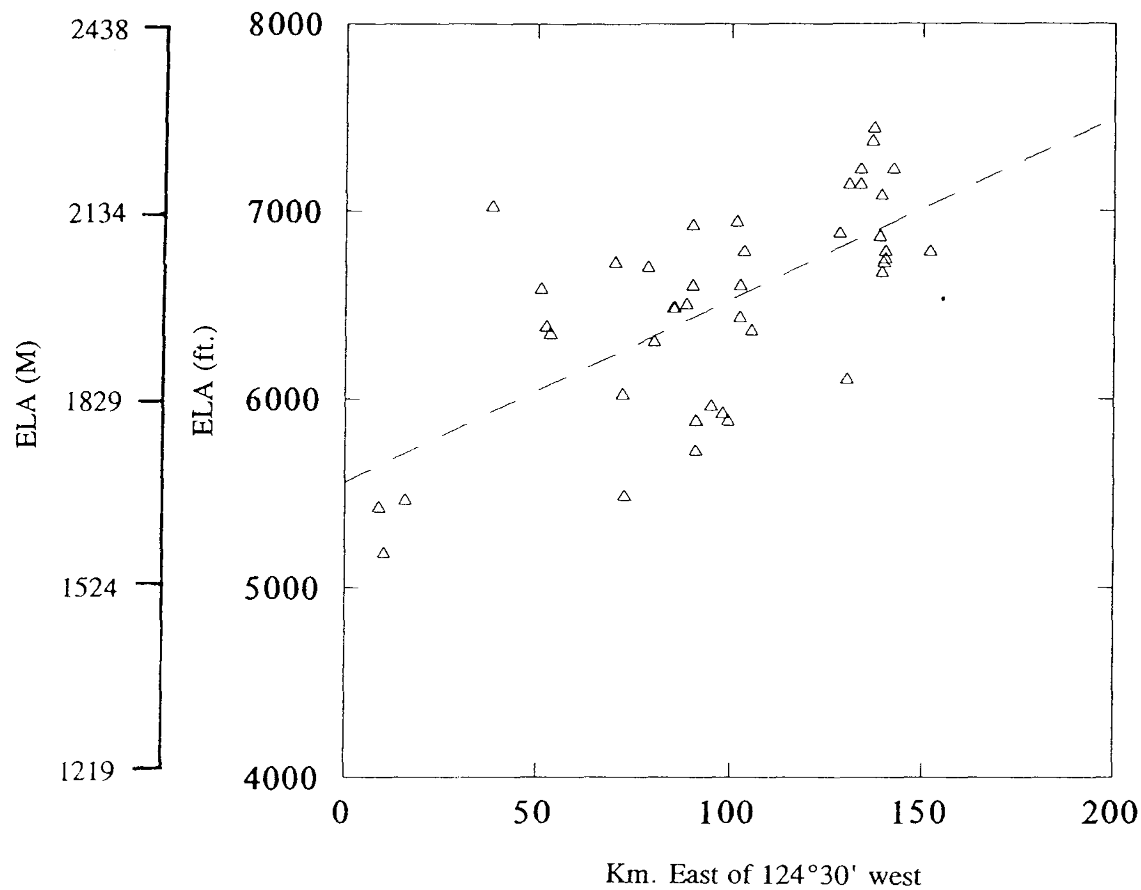
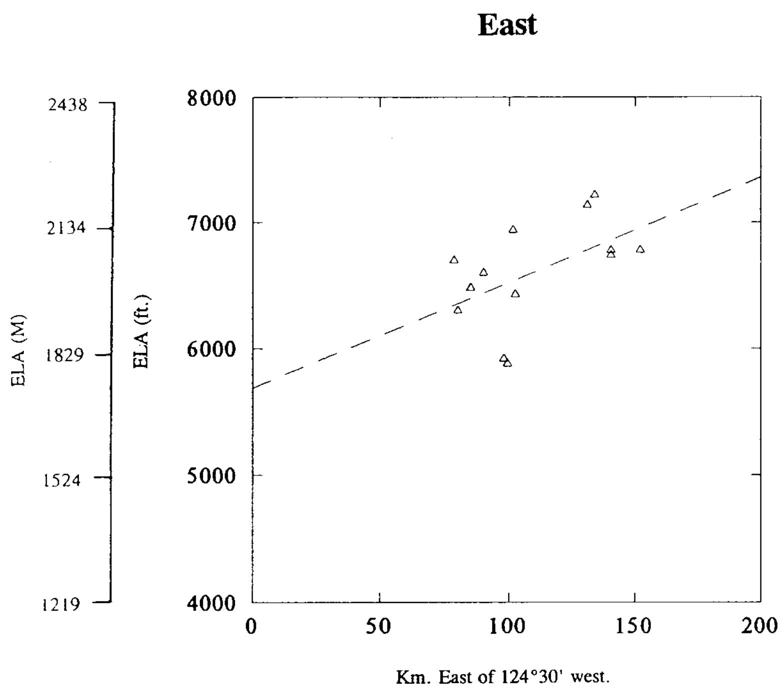
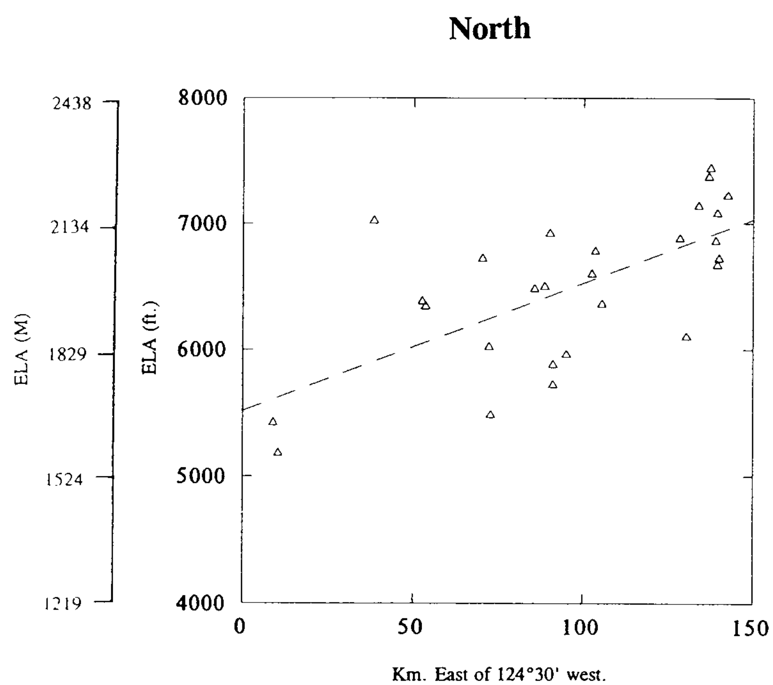


Figure 8.14 Modern ELA's for Glaciers with Evidence of LIA Extent, Derived by THAR Method - Effect of Aspect





8.3.1.2 Estimating Little Ice Age ELA's in the Southern Coast Mountains

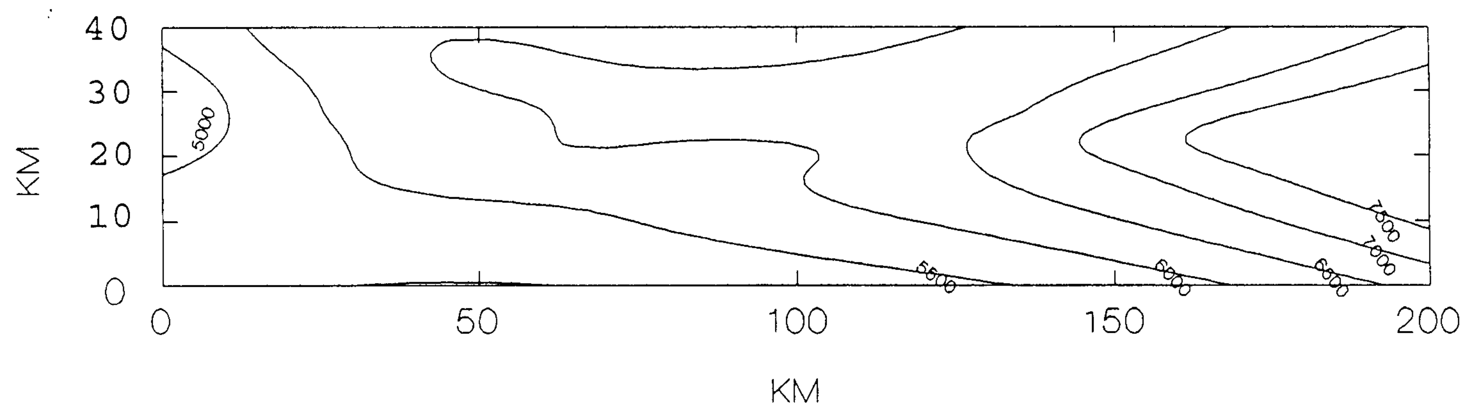
Estimation of LIA equilibrium lines requires that some morphological, biological or geological evidence of the extent of the former glacier be preserved. Further, interpretation of the estimates in terms of climate change requires knowledge of the time period during which the glacier maintained this former extent. Dating of physical evidence of glacial advance is difficult (See Figure 1.2). The principal periods of glacier advance in the Coast Mountains were the early 18th century and early 19th century. Dates on moraines are not available for sufficient numbers of glaciers to produce a transect of ELA's for a specific date. However the LIA advance was the most extensive Neoglacial phase of advance (Mathews, 1951; Ryder & Thomson, 1986). In addition they are relatively recent so that evidence of the advance is still fresh. Therefore where glaciers exhibit fresh morphological features representing a recent advance the assumption can be made that it represents the LIA extent of the ice. No firm date for the advance can be assigned but it is assumed that the extensive LIA advances identified date from the late 19th century cold period.

Air photograph coverage for the five map areas in the study was examined at the range of available scales: 1:10000 to 1:50000. Each glacier for which the modern ELA had been calculated was examined to search for the extent of the LIA advance. The two principle forms of evidence utilised were terminal moraines, and fresh trimlines. The outline of the LIA terminal zone of the glacier was then transferred to the topographic maps. Where topographic features on the map clearly identified the position of the moraine this was straightforward. In difficult cases the moraines were traced onto mylar, and then reduced to the correct scale using a photocopier and retraced onto the maps. The terminal moraines provide good evidence of the altitude of the snout of the former glacier. However evidence of the former headwall altitude of the ice is harder to obtain. Therefore for glaciers which showed no evidence of higher former ice levels the modern value was assumed. The estimates of LIA ELA may therefore be slightly low. However it is expected that this is not a major source of error since transfer of mass down glacier ensures that most of the glacier thickening during an advance occurs in the lower half of the glacier. The error is most likely less significant than error due to the use of the 100' contour interval for calculation.

The LIA ELA's calculated are plotted in Figure 8.15 and 8.16.

8.4 Results and Discussion

Figure 8.17 shows the best fit lines through the calculated ELA's for modern and LIA glaciers. Given the scatter in the data these lines are essentially parallel. The west-east gradient is determined by declining precipitation and increased summer temperatures as conditions become more continental inland. The parallelism of the modern and past ELA surfaces suggests that circulation patterns across the Coast Mountains were not subject to major changes during the LIA. In particular this parallelism supports the assumption that precipitation was close to modern levels during the LIA. The average ELA depression calculated from the mid-point of the plot (100km.) is 375 ft. (114 m.). Transforming this by the local lapse rate of 0.70°C./100 m. (Appendix 3) gives a temperature depression of 0.8°C. Assuming that LIA conditions were at least as wet as the present (Section 8.2), this figure may be interpreted as a maximum temperature depression for the LIA. Grove (1988), citing Porter (1981) and Burbank (1981), suggests that mean annual temperatures in the Western Cordillera rose around 1°C. between 1770 and 1850. It is uncertain whether temperatures interpreted from ELA fluctuation are associated with mean annual. or summer temperatures. Theoretically summer temperatures controlling ablation are the most significant thermal component of mass balance. However Porter (1977) showed that the best explanation of glaciation thresholds in the Cascade range, to the south of the current study area, was given by winter precipitation and *annual* temperature. Given the proximity of Porters field area



Elevations in Ft.

Origin 124°30' West, 50°15' North.

Metric Equivalents

Feet	Metres
4500	1219
5000	1371
5500	1524
6000	1676
6500	1829
7000	1981
7500	2134
8000	2438
8500	2591

Figure 8.15 Contour Plot of LIA ELA's Derived by THAR Method

Figure 8.16 LIA ELA's for Glaciers in the Coast Mountains, Derived by the THAR Method.

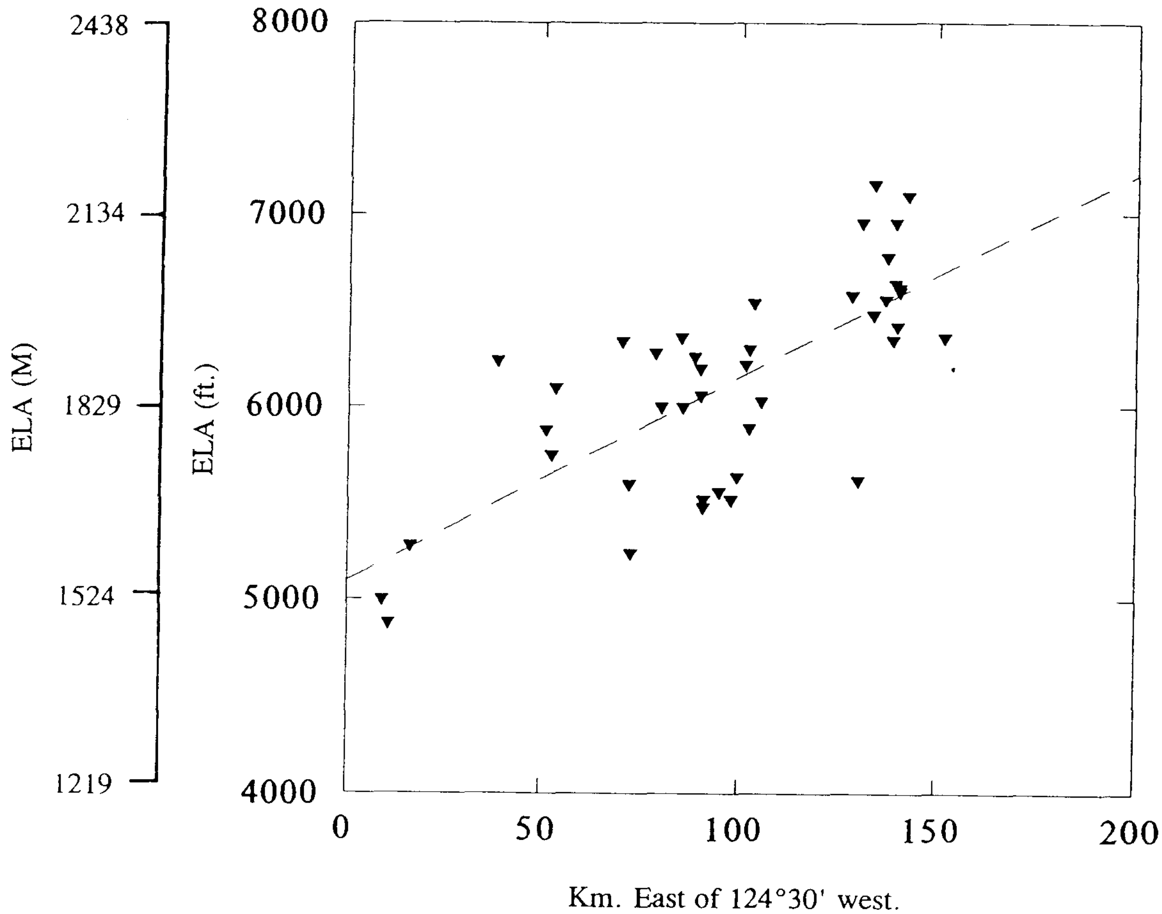
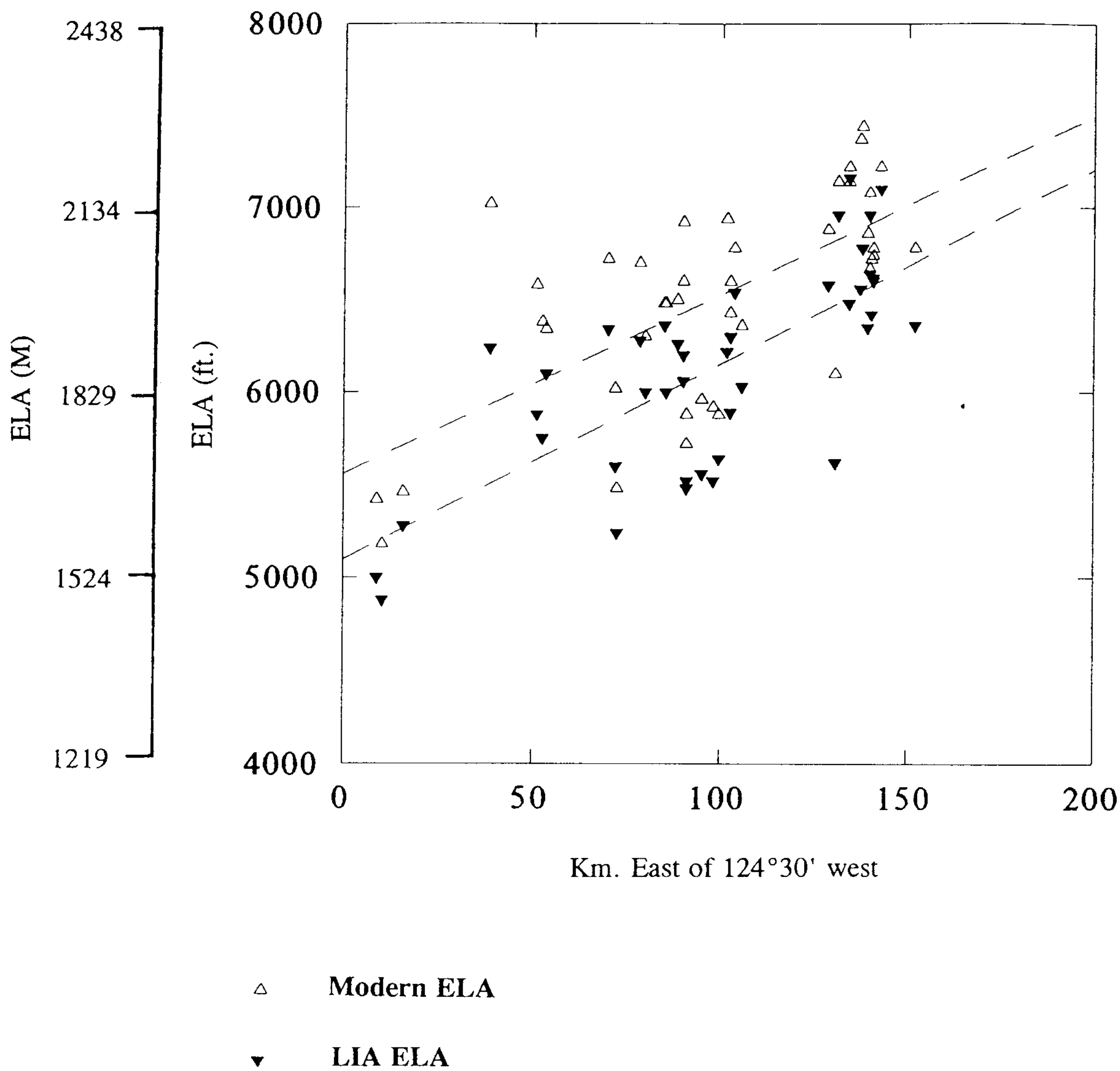


Figure 8.17 Modern and LIA ELA's across the Coast Mountains



to that of the current study it may be that the glacial fluctuations in the Coast Mountains are also correspond with changes in mean annual temperature.

From the discussion above it is clear that estimation of climatic variables from glacier fluctuations is a difficult procedure, fraught with uncertainty. However, methods adopted above produce a first order estimate of maximum LIA temperature depression in the Coast Mountains, which is consistent with published estimates from the region to the south.

The implications of the maximum decline of 0.8°C. in mean annual temperature for the LIA, inferred above are discussed in Chapter IX. This finding, along with those of the previous chapters are discussed in the light of previous knowledge of Neoglacial climates in the region.

Chapter IX Regional Context and Conclusions

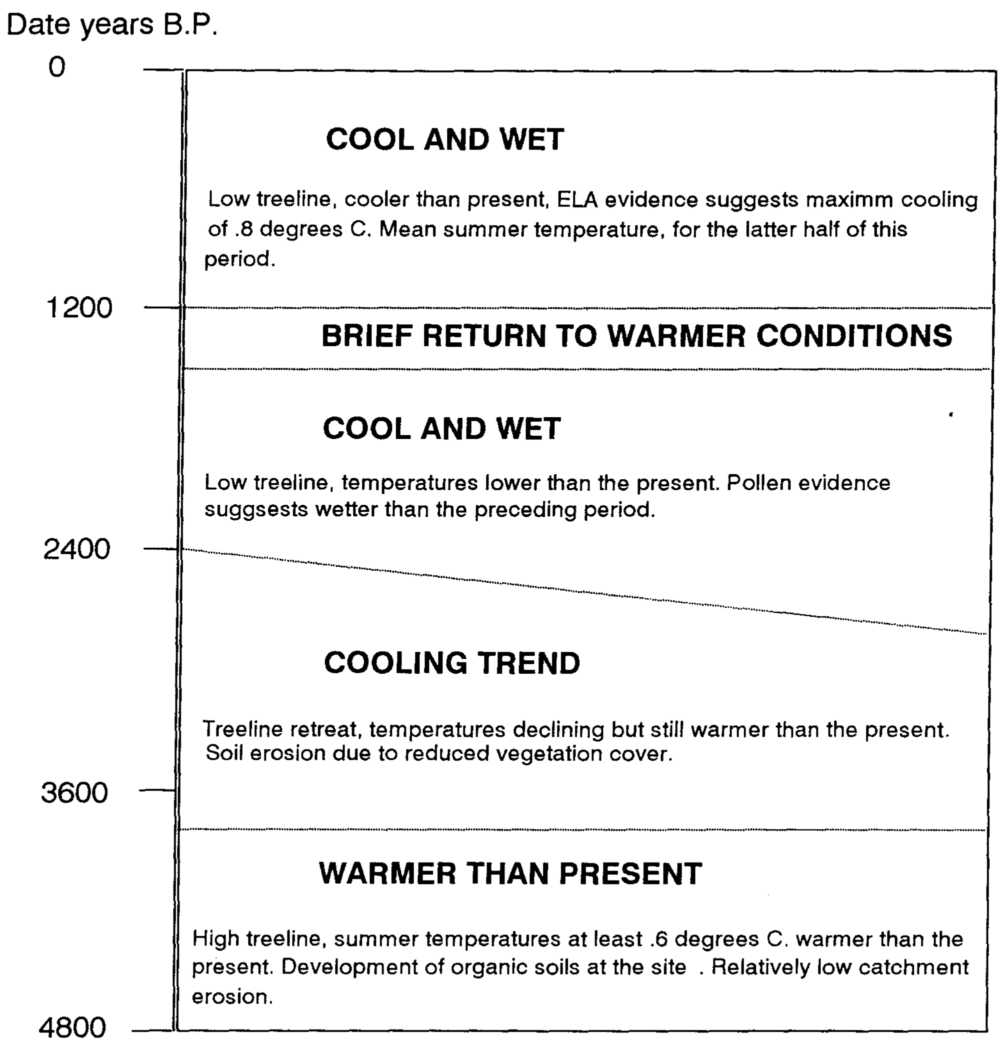
The objective of this thesis was to develop a more detailed understanding of the nature of Neoglacial climate changes in the southern Coast Mountains. In particular the thesis set out to produce a palynological record of Neoglacial climate change. Such a record is of interest for two reasons; it would allow direct comparison with palynological evidence of early Holocene climate, and it would provide a continuous record of Neoglacial climate change in contrast to the punctuated record provided by geomorphic evidence.

By careful site selection it has been possible to produce a pollen record which is apparently sensitive to the small climate changes of the Neoglacial. However even at such a sensitive site interpretation of percentage pollen data is not possible without reference to parallel lines of evidence. The pollen ratio technique (Maher, 1963) appears to allow quantification of the pollen record in terms of relatively small shifts in treeline elevation. Again the importance of the geographic location of sites for palaeoenvironmental reconstruction is emphasised. Careful site selection allowed interpretation of the quantified shifts in treeline elevation in terms of palaeotemperatures.

The substantive findings of the study, are summarised in Figure 9.1, and their implications for current understanding of Neoglacial climate in the region are discussed below:

The sedimentary evidence from Blowdown Lake covers the last 4800 years. Therefore, no evidence of climate changes assciated with the Garibaldi phase of ice advance (6000-5000 BP) was recovered. Pollen evidence from Blowdown Lake suggests that the period between 4800 and 3800 BP was marked by warm conditions with summer temperatures at least 0.6°C. above present. Between 3800 and 2500 BP

Figure 9.1 Summary of Findings



temperatures appear to have declined gradually to close to present values. Warm conditions at the site until 4000 BP present a problem of interpretation when viewed in a regional context. Mathewes and Heusser's (1981) calibration of the coastal pollen record suggests that the coolest period of the Holocene was from 6000-4000 BP (Although Heusser *et al.*, 1985, suggest that this was not the case in adjacent areas - Washington State, and southern Alaska). Similarly Ryder and Thompson suggest that there was no significant period of glacier contraction subsequent to expansion during the Garibaldi phase (6000-5000 BP) Two possible explanations for this apparent anomaly are advanced.

1) It was noted in Chapter 1 that the end of the Hypsithermal era in this region was a time transgressive event. On the southern coast cooling is apparent from 6000 BP onwards. In contrast warmer temperatures were maintained in the interior until 4500 BP. Since Blowdown Lake is positioned on the coast/interior ecotone it may be that it has experienced a similar climatic history to the interior. Therefore the decline in temperatures at the site at 3800 BP may represent the end of the Hypsithermal.

2) It may be that the complexity of the Hypsithermal/Neoglacial transition in southern B.C. results from the difficulty in separating temperature and precipitation effects, inherent in most proxy climate data sources. At Phair lake, to the southeast of the present study site Mathewes and King (1989) note correlations between lake level fluctuations and Neoglacial glacier advances. It may be that rising lake levels 5600 BP were due simply to increased precipitation, and not cooler temperatures as well. It is interesting to note that during the second phase of higher lake level identified by Mathewes and King (but not the first) molluscan macrofossil evidence suggests cooler water. The suggestion is therefore that climatic conditions in southern B.C. became wetter at 6000 BP but that cooling was delayed until about 4000 BP Therefore at the Blowdown Lake site, treeline, under temperature control, did not retreat until just after 4000 BP. This hypothesis may cast some light on the apparent complexity of the

Hypsithermal/Neoglacial transition in the region. Currently there is a strong climatic gradient from the coast to the interior. The coast experiences hyper maritime conditions whereas around 100 miles inland semi-desert conditions are found. If the change in climate at 6000 BP was to wetter conditions, the effect would be more apparent in coastal locations than inland because of the strong rainshadow effect of the coast mountains. This may provide an explanation for the correlation (admittedly weak) between continentality and the perceived end of the Hypsithermal. However, there is considerable variability in dates for the end of the Hypsithermal, even at sites in similar climatic zones. The hypothesis of changing precipitation may also provide a partial explanation of the apparent inconsistency of data on the timing of the close of the Hypsithermal. Precipitation in the coastal ranges exhibits a high degree of spatial variability. Local variability in space is associated with the rugged topography of the region. Current knowledge of the Hypsithermal/Neoglacial transition in B.C. is derived from relatively few sites which constitute a coarse sampling grid. Inconsistencies due to highly localised variations in precipitation should therefore be expected.

No evidence is available to distinguish between these alternatives. However it is interesting to note the possible explanation of the apparent complexity of the Hypsithermal/Neoglacial transition in the region provided by (3).

The pollen evidence has identified two periods of lower treeline, 2500-1500 BP and 1200 BP to present. These appear to correlate approximately with the Tiedemann and Late Neoglacial periods of ice advance (3300-1900 BP and 900-100 BP Ryder and Thompson, 1986). At Blowdown Lake calibration of pollen ratios for periods cooler than the present was not possible, due to lack of suitable sites for sampling modern pollen deposition. Therefore estimates of the magnitude of cooling in these intervals based on the pollen record were not possible. However evidence from estimates of ELA depression across the Coast Mountains suggests that during the maximum advance of the Little Ice Age temperature depressions were 0.8°C or less. Temperature depression would have been less than this figure if precipitation increased concurrently. The parallelism of modern and LIA equilibrium line altitude surfaces identified in this study suggests that precipitation levels were close to modern values. The estimate of 0.8°C maximum temperature depression during the late 19th century, is of similar magnitude to a 1°C rise in temperature from 1770-1850 A.D. estimated by Porter (1981) and Burbank (1981) for the Olympic peninsula (Grove 1988). Evidence that the last advances of the Little Ice Age were the most extensive of the Neoglacial (Ryder and Thompson, 1986) suggests that the 0.8°C temperature depression may be a maximum value for the whole Neoglacial.

Two reccomendations for future study arise from the discussion above. In order to avoid the difficulties of separating changes in precipitation from changes in temperature further work should concentrate on climatic proxies which are specific to temperature, or precipitation. This will allow testing of the hypothesis that temperature and precipitation changes at the end of the Hypsithermal were not synchronous. One possible approach would be further development of the techniques adopted in this study. Identification of treeline sites with abundant sites for sampling surface pollen and a long sedimentary record, may allow quantification of palaeotemperatures throughout the whole Neoglacial.

To summarise, this thesis has shown that with careful site selection, relatively small shifts in climate may be elucidated on the basis of the pollen record. The pollen ratio technique appears to allow quantification of these small changes. High treelines persisted at Blowdown Lake until 3800 B.P. Two periods of treeline lower than present have been identified and appear to correlate with the mid and late Neoglacial advances.

The quantification of Neoglacial climate provided by both palynological, and geomorphological evidence suggests that fluctuations of summer temperature on the order of 1.5°C have been experienced in the last 5000 years. Such natural small-scale variability has had significant and detectable effects on the nature of organic and clastic sedimentation in Blowdown lake. Recognition of such natural, rapid, short term changes in climate; and their effects on the landscapes and ecosystems of sensitive regions, are important to considerations of the nature of anthropogenic effects on the earth atmosphere system.

References

Adam, D.P. and West, G.J. (1983) Temperature and precipitation estimates through the last glacial cycle from Clear Lake, California, pollen data. *Science* **219**, 168-170.

Andrews, J.T. (1975) *Glacial Systems*. Duxbury Press, North Scituate, Mass.

Andrews, J.T., Carrara, P.E., King, F.B., and Stuckenrath, R. (1975) Holocene environmental changes in the alpine zone, northern San Juan Mountains, Colorado: evidence from bog stratigraphy and palynology. *Quat. Res.*, **5**, 173-197.

Antevs, E.V. (1938) Post pluvial climatic variations in the south west. *Am. Met. Soc. Bull.*, **19**, 190-193.

Arno, (1991) *Timberline: mountain and arctic forest frontiers*. Seattle, The Mountaineers.

Barry, R.G. (1992) *Mountain Weather and Climate* Routledge, London.

Beaudoin, A.B. (1986) Using pinus/picea ratios from the Wilcox Pass core, Jasper National Park, Alberta, to investigate Holocene timberline fluctuations. *Geog. Phys. Quat.* **XL(2)**, 145-152.

Behre, K.E. (1986) *Anthropogenic Indicators in Pollen Diagrams*. Balkema, Rotterdam.

Belcher R. and Ingram G. (1950) A rapid micro-combustion method for the determination of carbon and oxygen. *Anal. Chem. Acta.* **4,** 118-129.

Birks, H.J.B. and Birks, H.H. (1980) *Quaternary Palaeoecology,* Edward Arnold, London

Bradley, R.S. (1985) *Quaternary Paleoclimatology: Methods of paleoclimatic reconstruction.* Boston, Allen and Unwin.

Burbank, D.W. (1981) A chronology of late Holocene glacier fluctuations on Mount Rainier, Washington, U.S.A. *Arct. Alp. Res.* **13** 369-386.

Clapperton, C.M. Glacial geomorphology , Quaternary glacial sequence and palaeoclimatic inferences in Ecuadorian Andes. In Gardiner, V. (ed.) *International Geomorphology*, Part II 843-870. John Wiley and Sons, Chichester.

Colwell, R.N. (1951) The use of radioactive isotopes in determining spore distribution patterns. *Am. J. Bot.* **38**, 511-523.

Deevey, E.S. and Flint, R.F. (1957) Post glacial Hypsithermal interval. *Science*, **125,** 182-184.

Denton, G.H. and Karlen, W. (1973) Holocene climatic variations - their pattern and possible cause. *Quat. Res.* **3,** 155-205.

Denton , G.H. and Karlen, W. (1977) Holocene glacial and treeline variations in the White River Valley and Skolai Pass, Alaska and Yukon Territory. *Quat. Res.*, **7**, 63-111.

Deslosges, J.R. (1987) *Paleohydrology of the Bella Coola River Basin: An assessment of Environmental Reconstruction* Unpublished PhD thesis, Department of Geography, University of British Columbia, Vancouver, 363p.

Dunwiddie, P.W. (1985) **D**ichotomous key to conifer foliage in the Pacific Northwest. *Northwest Sci.* **59** 185-191.

Evans, I.S. (1990) Climatic effects on glacier distribution across the southern Coast Mountains, B.C., Canada. *Annals of Glaciol.* **14** 59-64.

Faegri, K. and Iverson, J. (1989) *Textbook of Pollen Analysis.* IVth edition by Faegri, K., Kaland, P.E. and Krzywinski, K. Wiley, London.

Fall, P.L. (1992) Pollen accumulation in a montane region of Colorado, U.S.A: a comparison of moss polsters, atmospheric traps, and natural basins. *Rev Palaeobot. and Palynol.* **72**, 169-197.

Frenguelli, G., Spieksama, F.M., Bricchi, M., Romano, B., Mincigrucci, A.H., Dankaart, N.W., and Ferranti, F. (1991) The influence of air temperature on the starting dates of the pollen season of Alnus and Populus. *Grana*, **30**, 196-200.

Frenzel, B. (1969) Floren und vegetationsgesshichte seit dem ende des Tertiars (Historische Geobotanik) *Fortsch Bot* **31,** 309-319

Gale, H.S. (1915) Salines in the Owens, Searles, and Panamint Basins, south eastern California. *U.S. Geol. Surv. Bull.*, **580,** 251-323.

Gallie, T.M. (1983) *Chemical denudation and hydrology near tree limit, Mountains, British Columbia.* Unpublished PhD thesis, Dept. of Geography, University of British Columbia, Vancouver, 262p.

Gallie, T.M. and Slaymaker, H.O. (1984) Variable solute sources and hydrological pathways in a coastal subalpine environment. In Burt, T.P. and Walling, D.E. (eds.) *Catchment Experiments in Fluvial Geomorphology*. 347-357, Geo Books, Norwich.

Gallie, T.M. and Slaymaker, H.O. (1985) **Hydrological controls on stream chemistry** *Proceedings of the 15th Hydrology Symposium of the National Research Council of Canada, Quebec City,* 288-306.

Grove, J.M., (1988) *The Little Ice Age.* London, Methuen.

Hall, S.A. (1990) Pollen deposition and vegetation in the southern Rocky Mountains and Southwest Plains, U.S.A. *Grana* **29**, 47-61.

Hallam, J.S., Edwards, J.N, Barnes, B. and Stuart, K.J. A late glacial Elk with associated barbed points from High furlong, Lancashire. *Proc. Prehist. Soc.* **39,** 100-128.

Hansen, H.P. (1940) Palaeoecology of two peat bogs in southwestern British Columbia *Am. J. of Bot.* **27,** 144-149.

Hansen, H.P. (1947) Climate versus fire and soil as factors in the post-glacial forest succession in the Puget Lowland of Washington. *Am.J. of Sci.* **245,** 265-286.

Hansen, H.P. (1950) Pollen analysis of three bogs on Vancouver Island, Canada. *Ecology,* **38,** 270-276.

Hansen, H.P. (1955) Post-glacial forests in south-central, and central British Columbia. *Am. J. of Sci.* **253 (11)** 640-658.

Hebda, R.J. (1982) Postglacial history of grasslands of southern British Columbia. in Nicholson, A.C., McClean, A., and Baker, T.E. eds. *Grassland ecology and Classification: symposium proceedings.* British Columbia Ministry of Forests.

Heusser, C.J. (1960) *Late Pleistocene environments of North Pacific North America.* Am. Geog. Soc. Special Publication **35.**

Heusser, C.J. (1973) Modern pollen spectra from Mount Rainier, Washington. *Northwest Sci.* **47(1),** 1-9.

Heusser, C.J., Heusser, L.E., and Peteet, D.M. (1985) Late Quaternary climate change on the American North Pacific coast. *Nature* **315** 485-486.

Heusser, C.J., Heusser, L.E., and Streeter, S.S. (1980) Quaternary temperatures and precipitation for the northwest coast of North America. *Nature,* **286,** 702-704.

Hitchcock, C.L. and Cronquist A. (1973) *Flora of the Pacific Northwest* University of Washington Press, Seattle. 730p.

Jacobson, G.L. and Bradshaw, R.H.W. (1981) The selection of sites for paleovegetational studies. *Quat. Res.*, **16**, 80-96

Jones, P.S.A. (1988) *Postglacial chronology of large earthflows in south-central British Columbia.* Unpublished Ph.D. thesis , Dept. of Geography, University of British Columbia, Vancouver, 446 p.

Kearney, M.S. (1983) Modern pollen deposition in the Athabasca Valley, Jasper National Park. *Bot. Gaz.* **144(3)**, 450-459.

Kearney, M.S. and Luckman, B.H. (1983) Holocene timberline fluctuations in Jasper National Park, Alberta. *Science* **221**, 261-263.

LaMarche, V.C. (1973) Holocene climatic variations inferred from treeline fluctuations in the White Mountains, California. *Quat. Res.,* **3**, 632-660.

Leonard, E.M. (1984) Late Pleistocene equilibrium-line altitudes and modern snow accumulation patterns, San Juan Mountains, Colorado, U.S.A. *Arct. Alp. Res.* **16 (1)**, 65-76.

Livingstone, D.A. (1955) A lightweight piston sampler for lake deposits. *Ecology* **36,** 137-9.

Lowe, J.J. and Walker, M.J.C. (1984) *Reconstructing Quaternary Environments* Longman, Harlow. 389 p.

Luckman, B.H. (1986) Reconstruction of Little Ice Age events in the Canadian Rocky Mountains. *Geog. Phys. Quat.* **XL(1)**, 17-28.

Luckman, B.H., Holdsworth, G. and Osborn, G.D. (1993) Neoglacial fluctuations in the Canadian Rockies. *Quat. Res.* **39** 144-153.

Mackereth, F.J.H. (1966) Some chemical observations on post-glacial lake sediments. *Phil. Trans. R. Soc.* **Series B 250** 165-213.

Maher, L.J. (1963) Pollen analyses of surface materials from the southern San Juan Mountains, Colorado. *Geol. Soc. Am. Bull.* **74**, 1485-1504.

Maher, L.J. (1972) Absolute pollen diagram of Redrock Lake, Boulder County, Colorado. *Quat. Res.,* **2**, 531-553.

Mann D.H. (1986) Reliability of a fjord glacier fluctuations for palaeoclimatic reconstructions *Quat. Res.,* **25** 10-24.

Markgraf, V. (1980) Pollen dispersal in a mountain area. *Grana* **19**, 127-146.

Mathes, F.E., (1939) Why we should measure glaciers. *Sierra Club Bulletin*, **20,** 20-27.

Mathewes, R.W. (1973) A Palynological study of postglacial vegetation changes in the University Research Forest, southwestern British Columbia. *Can. J. Bot.* **51,** 2085-2103.

Mathewes, R.W. (1985) Paleobotanical evidence for climatic change in southern British Columbia during Late-Glacial and Holocene time. in Harington C.R. ed. Climatic change in Canada 5: Critical periods in the Quaternary climatic history of Northern North America. *Syllogeus,* **55,** 397-422.

Mathewes, R.W. (1988) Pollen spectra from lake surface sediments in the Swiss Alps. *Arct. Alp. Res.* **20(3)**, 310-324.

Mathewes, R.W. and Heusser, L.E. (1981) A 12000 year palynological record of temperature and precipitation trends in southwestern British Columbia. *Can. J. Bot.* **59,** 707-710.

Mathewes, R.W. and King, M. (1989) Holocene vegetation, climate, and lake level changes, in the Interior Douglas Fir biogeoclimatic zone of British Columbia. *Can. J. Earth Sci.* **26,** 1811-1825.

Mathewes, R.W. and Rouse., G.E. (1975) Palynology and Palaeoecology of postglacial sediments from the lower Fraser River Canyon of British Columbia. *Can. J. Earth Sci.* **19,** 2114-2125.

Mathewes, R.W. and Westgate, J.A. (1980) Bridge River tephra: revised distribution and significance for detecting old carbon errors in radiocarbon dates of limnic sediments in southern British Columbia. *Can. J. Earth. Sci.* **17**, 1454-1461.

Mathews, W.H. (1951) Historic and prehistoric fluctuations of alpine glaciers in the Mount Garibaldi map-area, south-western British Columbia. *J. Geol.* **59,** 357-387.

Meier, M.F. and Post, A. (1962) Recent variations in mass net budgets of glaciers in western North America. *IUGG, Int. Ass. of Sci. Hydr., Comm. of Snow and Ice,* Publication no. **58** 63-77

Meirding, T.C. (1982) Late Pleistocene glacial equilibrium-line in the Colorado Front Range: a comparison of Methods. *Quat.Res.* **18,** 289-310.

Minore, D. (1979) *Comparative autecological characteristics of northwestern tree species: a literature review.* Pacific Northwest Forest and Range Experiment Station, U.S. Dept. of Agriculture, Forest Service, General Technical Report PNW-87

Moore P.D. and Webb J.A. (1978) *An illustrated guide to pollen analysis*, Hodder and Stoughton, London.

Oldfield, F. and Robinson, S.G. (1985) Geomagnetism and palaeoclimate. In *The Climatic Scene* Tooley, M.J. and Sheil, G. (eds) George Allen and Unwin, London.

Ostrem, G. (1966) The height of the glaciation limit in southern British Columbia and Alberta. *Geografiska Annaler* **46 A** 127-138.

Ostrem, G. (1973) The transient snowline and glacier mass balance in southern British Columbia and Alberta, Canada. *Geografiska Annaler* **55 A** 93-106.

Owens, P.N. (1990) *Lake sediment based sediment yields and erosion rates in the Coast Mountains, British Columbia.* Unpublished MSc thesis, Dept. of Geography, University of British Columbia, Vancouver, 138 p.

Patterson W.A., Edwards, K.J., and McGuire, D.J. (1987) Microscopic charcoal as a fossil indicator of fire. *Quat. Sci. Rev.* 6, **3-23.**

Pennington, W. and Lishman, J.P. (1970) Iodine in lake sediments in northern England and Scotland *Biol. Rev. 46, 279-313.*

Porter, S.C. (1975) Equilibrium-line altitudes of late Quaternary glaciers in the Southern Alps, New Zealand. *Quat.Res.* **5** 27-45.

Porter, S.C. (1977) Present and past glaciation threshold in the Cascade Range, Washington, U.S.A.: topographic and climatic controls, and palaeoclimatic implications. *J. Glaciol.*, **18,** 101-116.

Porter, S.C. (1981) Lichenometric studies in the Cascade range of Washington: establishment of *Rhizocarpon geographicum* growth curves at Mount Rainier. *Arct. Alp. Res* **13** 11-23.

Porter, S.C. (1981) Glaciological evidence of Holocene climatic change. In Wigley, T.M.L., Ingram, M.J. and Farmer, G. (eds.) *Climate and History: Studies in past climates and their impact on man.* 82-110, Cambridge University Press, Cambridge.

Porter, S.C., Pierce, K.L., Hamilton T.D. (1986) Late Wisconsin mountain glaciation in the Western United States. In Porter, S.C. (ed.) *Late Quaternary environments of the United States,* 71-114, Longman Group Ltd., London

Porter, S.C. and Denton, G.H. (1967) Chronology of Neoglaciation in the North American Cordillera. *Am. J. Sci.* **265,** 177-210.

Reynaud, L. (1980) Can the linear balance model be extended to the whole Alps?, in Muller, F. (ed.) *World Glacier Inventory. Proceedings of the Riederalp Workshop, 1978,* International Association For Hydrological Sciences Publ. no. **126** 273-284

Rummery, T.A. (1983) The use of magnetic measurement in interpreting the fire histories of drainage basins. *Hydrobiologia,* **103,** 53-58.

Rummery, T.A., Bloemendal, J., Dearing, J., Oldfield, F. (1979) The persistence of fire-induced magnetic oxides in soils and lake sediments. *Ann. Geophys.* **35,** 103-107.

Ryder, J.M. and Thomson, B. (1986) Neoglaciation in the southern Coast Mountains of British Columbia: Chronology prior to the late Neoglacial maximum. *Can. J. Earth Sci.* **23,** 273-287.

Ryder, J.M., Thomson, B., and Alley, N.F. (1981) *Neoglacial Chronology For the Southern Coast Mountains of British Columbia* Unpublished report.

Sarna-Wojcicki, A.M., Champion, D.E. and Davis, J.O., (1983) Holocene volcanism in the conterminous United States and the role of silicic volcanic ash deposits in correlation of latest-Pleistocene and Holocene deposits. In Wright H.E. (Ed), *Late-Quaternary Environments of the United States. Volume 2: The Holocene.* University of Minnesota Press, Minneapolis, 52-77.

Schmok, G. (1986) *Sedimentology and Chronology of Neoglacial Lake Alsek, Yukon Territory.* Unpublished MSc thesis, Dept. of Geography, University of British Columbia.

Shackleton, N.J and Opdyke, N.D. (1973) Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-128: oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year scale *Quat. Res.*, **3,** 39-55.

Sissons, J.B. (1980) The Loch Lomond Advance in the Lake District, northern England, *Trans. R. Soc. Edinb. Earth Sci.* **71**, 13-27

Solomon, A.M. and Silkworth, A.B. (1986) Spatial patterns of atmospheric pollen transport in a montane region. *Quat. Res.*, **25**, 150-162.

Souch, C.J. (1984) *Holocene lacustrine sedimentation in a small subalpine watershed.* Unpublished MSc thesis, Dept. of Geography, University of British Columbia. Vancouver. 154 p.

Souch, C.J. (1990) *Lake sediments as records of palaeoenvironmental change, Kwoiek Creek, Coast Mountains, B.C.* Unpublished Ph.D. thesis, Dept. of Geography, University of British Columbia, Vancouver. 214 p.

Stuiver, M., Deevey, E.S., and Gralenski, L.J. (1960) Yale natural radiocarbon measurements V. *Am. J. Sci. Radiocarbon Supplement*, **2**, 49-61.

Sugden, D.E. and John B.S. (1976) *Glaciers and Landscape*. London, Edward Arnold.

Tauber, H. (1965) Differential pollen dispersion and the interpretation of pollen diagrams. *Dan. Geol. Unders.* **2:69**, 69 p.

Thomson, R. and Oldfield, F. (1986) *Environmental Magnetism*. Allen and Unwin, London.

Thompson, R., Battarbee, R.W., O'Sullivan, P.E., and Oldfield, F. (1975) Magnetic susceptibility of lake sediments. *Limnol. Oceanogr.* **20,** 687-698.

Torsnes, I., Rye, N., and Nesje, A. (1993) Modern and Little Ice Age equilibrium-line altitudes on outlet valley in glaciers from Jostedalsbresn, Western Norway: An evaluation of different approaches to their calculation. *Arct. Alp. Res.* **25 (2)**, 106-116.

Tranquilini, W. (1979) *Physiological Ecology of the Alpine Timberline*. Springer Verlag, Berlin.

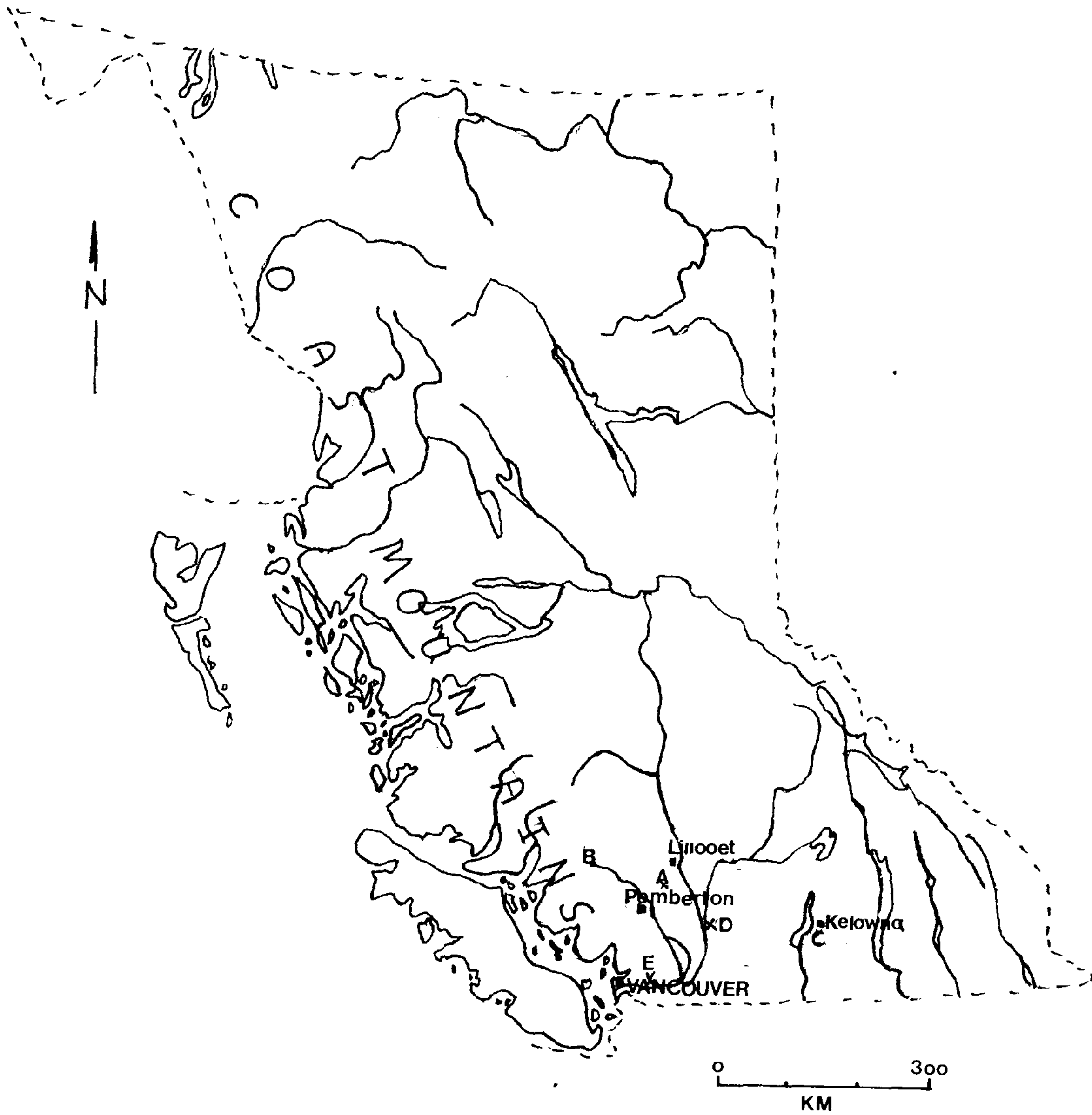
Troels-Smith, J. (1955) Karakterisering af lose jordater. Characterisation of unconsolidated sediments. *Danm. Geol. Unders. Series IV* **3(10)** 1-73.

Wainman, N. and Mathewes, R.W. (1990) Distribution of plant macroremains in surface sediments of Marion lake, southwestern British Columbia. *Can. J. Bot.* **68(2)** 364-373.

Woodsworth, J. (1977) *Geology of the Pemberton (92-J) map area.* Geological society of Canada open file report 482, Ottawa, 1 map sheet.

Appendix 1

Figure A1.1 Map of British Columbia Showing Sites Mentioned In Chapter 1



A = Blowdown Lake

B = Lillooet Ice Field

C = Kelowna Bog

D = Fraser Canyon

E = Marion Lake

Appendix 2

This appendix presents in graphical format the visual stratigraphy of all the cores from Blowdown Lake. The graphical key is presented below.

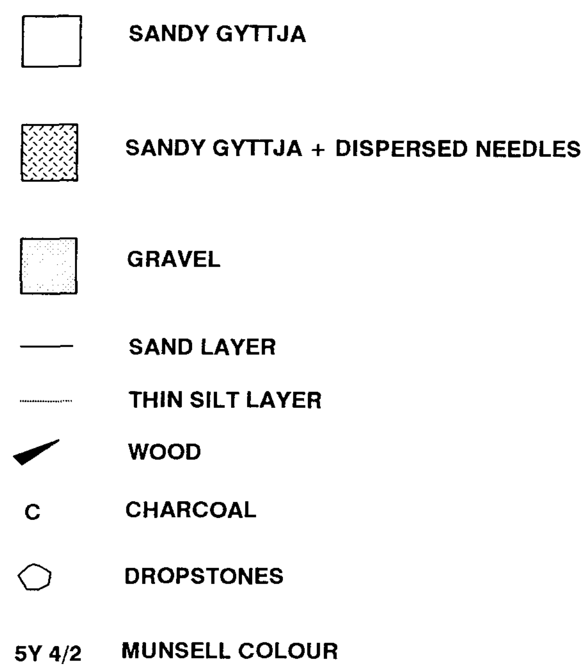


Figure A2.1 Visual Stratigraphy of Blowdown Lake Cores

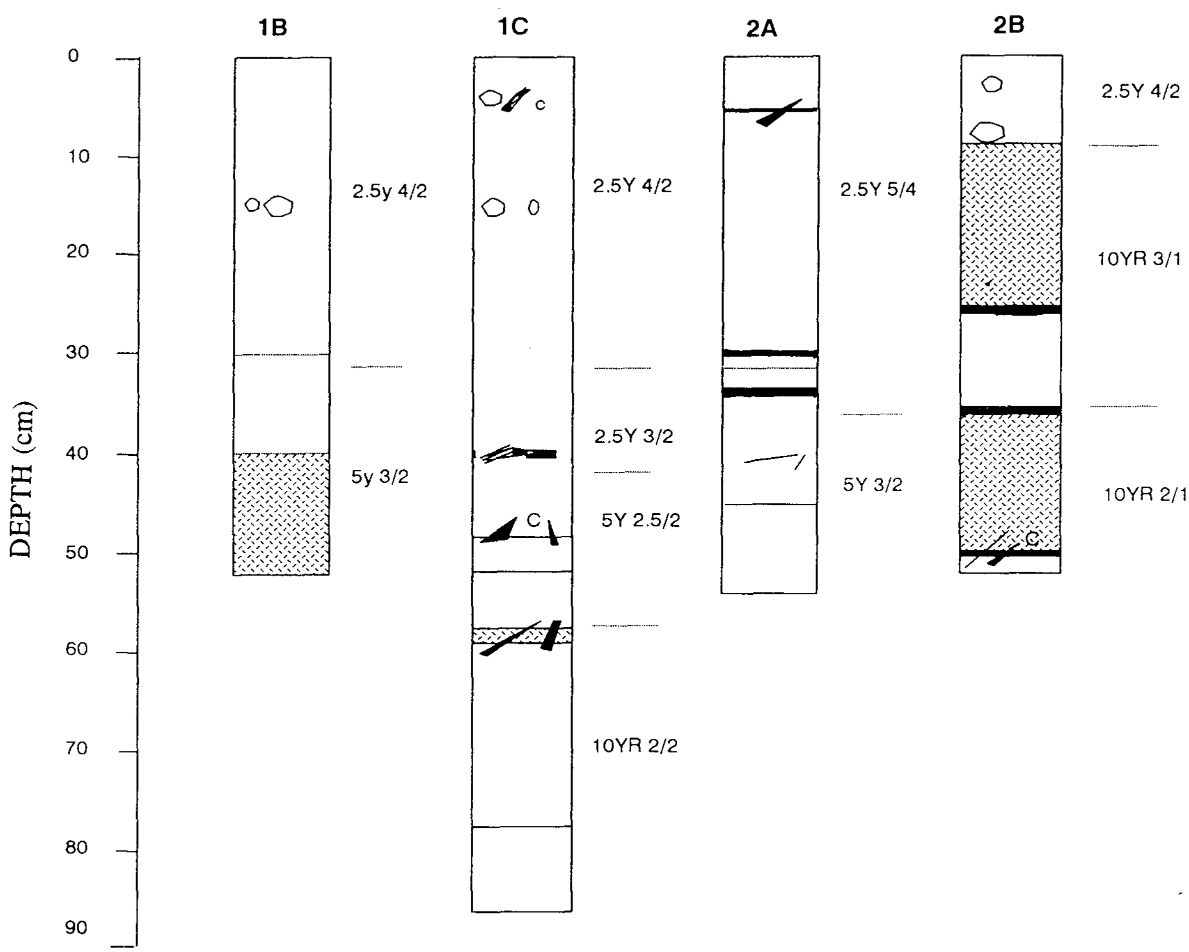
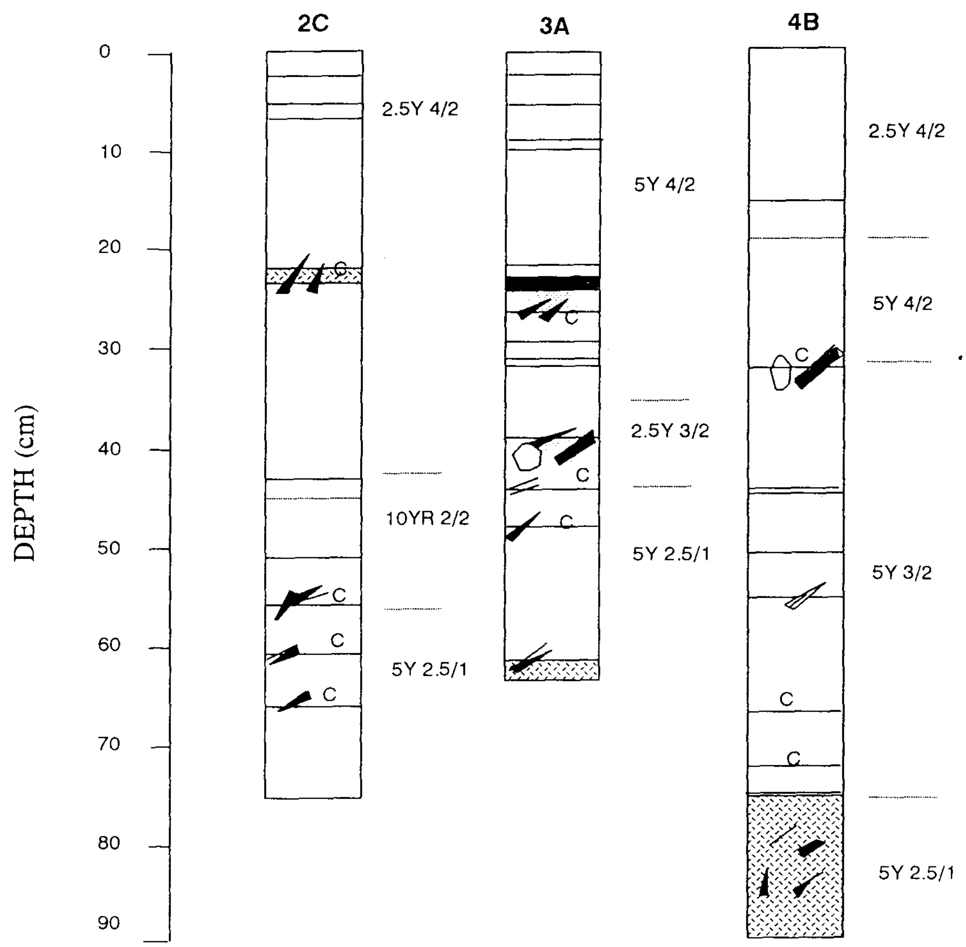


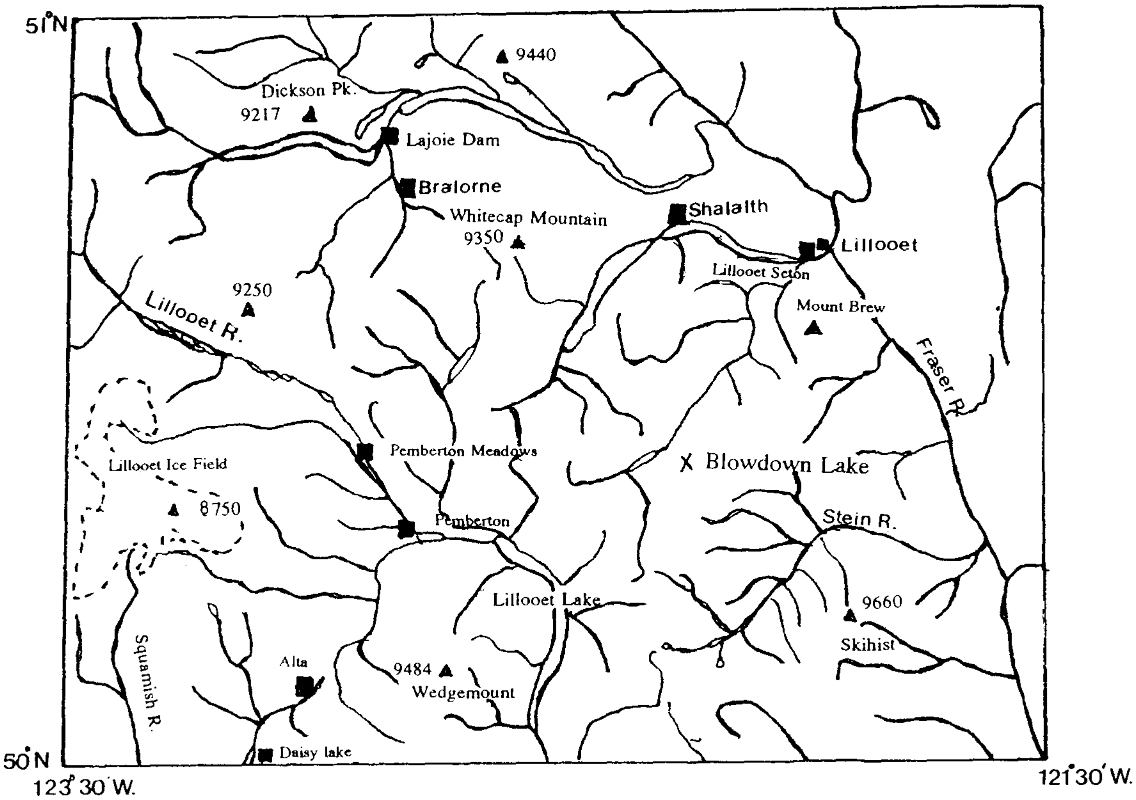
Figure A2.1 (continued)



Appendix 3

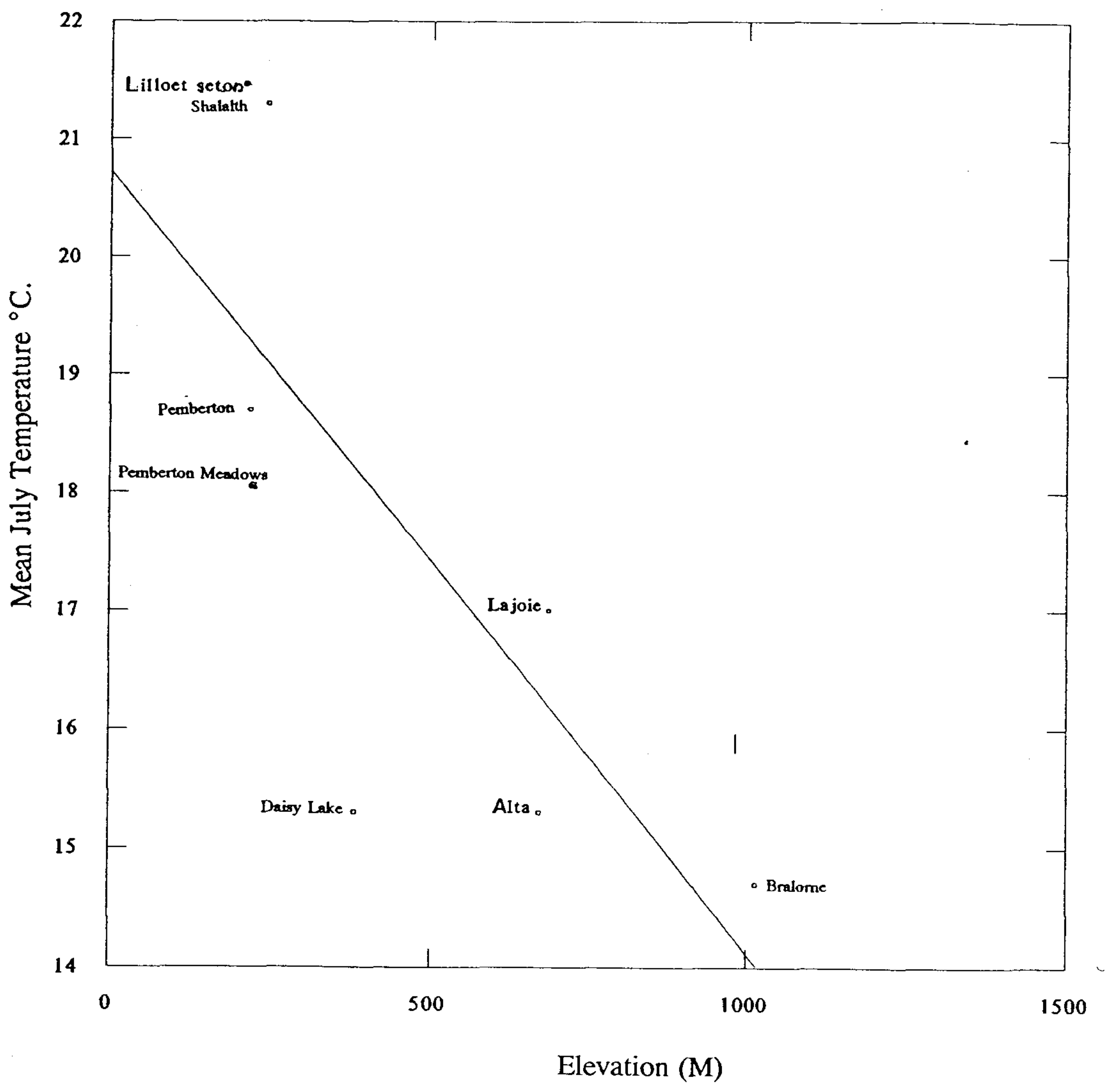
In order to calibrate tree line shifts in terms of temperature an estimate of the mean local lapse rate is required. The lapse rate of interest is a mountain lapse rate, i.e. change in temperature upslope, rather than a free air lapse rate. Since this thesis is concerned with estimating summer temperatures this becomes the summer mountain lapse rate. In order to estimate this rate mean July temperatures (1951-1981 climatic normal) for local climate stations (Figure A3.1) were plotted against elevation (Figure A3.2). The slope of the best fit line through these points is 0.007. 0.7°C./100 m is therefore taken as an approximation to the local mean summer mountain lapse rate. It is clear from A3.1 that climate stations in the region are sparse, and confined to valley floors and therefore unrepresentative of topographic variability. In addition the climate data plotted in A3.2 are from varying climatic regimes. It is therefore unsurprising that the standard error of the best fit line in A3.2 is large. This line is only an approximation to the real mountain lapse rate. However it does at least provide some local grounding for the lapse rate figure used for estimates of Neoglacial thermal change in chapters VI and VIII. The value of 0.7 is close to the often cited mean lapse rate figure of 0.65°C/100m.

Figure A3.1 Location of Local Climate Stations



■ Climate Stations

Figure A3.2 Plot of Mean July Temperature against Station Elevation



$Y = -.007x + 20.725$

$R^2 = .571$

Std. Err. Est. = 1.87