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# Olympia Interstadial: vegetation, landscape history, and paleoclimatic implications of a mid-Wisconsinan (MIS3) nonglacial sequence from southwest British Columbia, Canada

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**Abstract :**

Lithostratigraphic, C-14, and palynologic analyses of peat and silty peat at three nearby sites reveal a 25 000 year vegetation and climate history of the Olympia Interstade for the Fraser Lowland, British Columbia, 300 km within the southern limit of the Cordilleran Ice Sheet. At Lynn Valley, Polypodiaceae fern spores and nonarboreal pollen dominate >47.8 C-14 ka BP, reflecting unstable and cold landscapes. A Pinus-Poaceae zone follows, representing pine parkland and cool dry climate. Fluctuating values of Picea and Tsuga mertensiana pollen at Lynn and Seymour valleys and Port Moody characterize most of the Olympia Interstade during local peat deposition in Cyperaceae and Myrica wetlands until about 26.7 C-14 ka BP under a cool and moist climate. A brief Pinus - Tsuga heterophylla zone at Lynn Valley 44-39 C-14 ka BP suggests a climatic optimum. A Poaceae-Artemisia assemblage and deposition of silty sand after 26.7 C-14 ka BP reflect cooling and drying after which a unique Lycopodium assemblage at Lynn Valley suggests cold arid climate and Fraser Glaciation onset. These sequences have no progression to vegetation typical of warm, interglacial, Holocene-like climates, indicating an interstadial not an interglacial interval. Correlation with vegetation changes elsewhere in western North America suggests that the Olympia Interstade started about similar to 52 C-14 ka BP (similar to 57 cal ka BP) and ended about 26 C-14 ka BP (30 cal ka BP).

**Keywords** : Pollen, Stratigraphy, Quaternary, Radiocarbon dates, Paleoecology

## Introduction

The Olympia nonglacial interval preceded major Late Wisconsin ice advances and has long been of interest in northwest North America (Heusser 1977, Clague and MacDonald 1989, Mathewes 1991, Cosma et al. 2008). Viewed by some as an interstadial, and by others as a full interglacial interval, it is of particular interest because it spans a time of rapid hemispheric climatic changes including the run-up to the last glacial maximum of the Cordilleran Ice Sheet (CIS; Jimenez-Moreno et al. 2010). There is interest in the pre-conditions and dynamics of rapid climatic changes in the context of global climate change (Clark et al. 2009). Of particular interest are the succession and rates of change of plant communities in response to climatic changes and the variation of such responses on local, regional and hemispheric scales (Grigg et al. 2001; Jimenez-Moreno et al. 2010).

The concept of a long non-glacial "Olympia" interval that immediately preceded the Fraser Glaciation (marine isotope stage 2, MIS2) has long been recognized in northwest North America (Armstrong et al. 1965; Clague 1978). It was first considered an interglaciation (Armstrong et al. 1965) more or less equivalent to MIS3. Later it was changed to the more general Olympia nonglacial interval by Hansen and Easterbrook (1974) and Armstrong and Clague (1977) because of differing paleoclimatic interpretations between southwest and southern interior British Columbia (BC) (e.g. Fulton 1971; Clague 1976; Alley et al. 1986) and northwest Washington State (e.g. Hansen and Easterbrook 1974; Heusser 1977). Clague (1978) reviewed the paleoclimate controversy associated with this interval and concluded that the Olympia was a lengthy non-glacial episode characterized by sharply fluctuating, but generally cool climate. Based on palynologic evidence from southern Vancouver Island, and comparisons with records in northwest Washington State, Alley (1979) reverted to Olympia Interglaciation claiming that the climate during the interval was similar to present. In contrast, oxygen isotope data from a speleothem sampled in Cascade Cave on southern Vancouver Island indicated to Gascoyne et al. (1981) that Olympia climate was cooler than present throughout. Clague (1981) compiled radiocarbon ages for BC and concluded that during the Olympia temperatures were at times similar to, and at times cooler than present. In a subsequent review Clague and MacDonald (1989) concluded that Olympia climate was variable but generally cooler than present. Recently Cosma et al. (2008:941) referred to the interval formally as the Olympia Interstade, while discussing it in terms of the Olympia non-glacial interval (Cosma et al. 2008:951). In this paper we use the term Olympia Interstade based on our new data from the most complete dated terrestrial sequences of this nonglacial episode in southwest BC.

We report on a ~25 000-year continuous record from the last major non-glacial episode in southwest BC, from sites within 300 km the southern limit of the last CIS. Using stratigraphic profiles and pollen and spore analyses of four separate contemporaneous sites we reconstruct the sequence of vegetation and infer climate and the character of the landscape from more than 48 $^{14}$C ka BP to about 26.7 $^{14}$C ka BP. We use the radiocarbon time scale for primary regional correlation and description because previous key treatments, such as those of Heusser (1977), do so. Moreover, calibration of the radiocarbon timescale in the older parts of the Olympia interval is not well established. We compare the reconstructed plant communities to modern vegetation and, in combination with the sedimentary record, reconstruct the history of the landscape. Next we compare and contrast the sequence to adjacent records within and outside the glacial limit, especially those of Washington and Oregon states. We correlate Fraser Lowland events with regional chronology and biome changes elsewhere in western North America. In so doing we resolve the question of whether or not the Olympia was an interglacial, and define its length, and start and end times. We also provide a continuous, well-dated Olympia palynostratigraphic framework for correlating and interpreting other sequences. This is the highest resolution sequence studied within CIS limits based on multiple sites, and complements long records of the west side of the Olympic Peninsula (Heusser 1977, Heusser et al. 1999) and those south of CIS limits (Grigg et al. 2001, Grigg and Whitlock 2002, Jimenez-Moreno et al. 2010).

## Study area and sites

### *Regional ecology and paleoecology*

The Fraser Lowland currently supports temperate conifer rainforests of the Coastal Western Hemlock Biogeoclimatic zone developed largely on podsols (CWH; Meidinger and

Pojar 1991) in which *Tsuga heterophylla* (western hemlock) and *Thuja plicata* (western red cedar) dominate and *Abies* (mostly *Abies amabilis* (amabilis fir)) occurs abundantly in moister climates. *Pseudotsuga menziesii* (Douglas-fir) grows in the warmest and driest areas, and *Alnus rubra* (red alder) is characteristic of disturbed sites. At elevations of 1000 to 1800 m CWH forests are replaced by Mountain Hemlock (MH) biogeoclimatic zone forests, the lower elevations of which are dominated by *Tsuga mertensiana* (mountain hemlock), *Chamaecyparis nootkatensis* (yellow cedar) and *Abies amabilis* stands. The upper elevations of this zone comprise parkland with open shrubby communities of *Phyllodoce* spp. and *Cassiope* spp. (heathers) and herbs. High elevation Alpine Tundra (AT) consists of heath, and herbaceous vegetation and rock outcrops. Eastward under more continental climates, the MH zone is replaced by forests and parkland of the Engelmann Spruce-Subalpine fir (ESSF) biogeoclimatic zone. *Picea engelmannii* (Engelmann spruce) and *Abies lasiocarpa* (subalpine fir) dominate the canopy.

Broadly speaking, each of these major modern ecosystems has been observed in Holocene, Late-glacial and Late Pleistocene non-glacial pollen records in the Fraser Lowland (Hebda and Whitlock 1997). Conifer forests characterize the Late Pleistocene and Holocene vegetation in the region (e.g., Mathewes 1973; Hebda 1995; Pellatt et al. 2002). Immediately before the Vashon glacial advance (maximum ~14.5 $^{14}$C ka; Hicock and Armstrong 1985, Clague and Ward 2011) at ~17.5 BP, ESSF forest and parkland occurred during the brief Port Moody Interstade (Hicock et al. 1982, 1999; Miller et al. 1985; Hicock and Lian 1995; Lian et al. 2001). During the early part of the Fraser Glaciation tundra-like vegetation occurred in brief intervals at Point Grey and on adjacent Vancouver Island (Mathewes 1979; Alley 1979). Olympia plant assemblages from Lynn valley (Armstrong et al. 1985 provided only a summary diagram) include forest and open vegetation. Outside the limits of the CIS, in Washington and Oregon states, non-arboreal tundra-like vegetation is recorded during the Fraser Glaciation (Whitlock 1992; Hebda and Whitlock 1997) and at the beginning of the Olympia interval, with coniferous forests during the rest of the Olympia (Heusser et al. 1999).

***Study Sites***

The four natural exposures in this study occur near the northwest boundary of the Fraser Lowland, at the south edge of the Coast Mountains, approximately 300 km north of the southern limit of the last CIS (Fig. 1). The Lynn Valley sections occur near the mouth of Lynn Creek: one (LW) on the west side the creek, and the other (LE) 185 m away on the east side. The base of both sections is at ~116 metres above sea level (m asl). The Seymour valley (SV) section is in the adjacent valley on the west bank of Seymour River, elevation ~90 m asl, 1.7 km northwest of the Lynn Creek sites. The Port Moody (PM) section occurs about 12 km southeast of the Lynn and Seymour sites at ~50 m asl.



The lower reaches of Lynn and Seymour valleys have dissected valley-fills up to 100 m thick, most of which was deposited during the Fraser Glaciation. The sediment fill in Lynn valley has not been thoroughly described. However, in adjacent Seymour valley it includes glaciofluvial and glaciolacustrine sediments, and till, associated with two ice advances which occurred during the Coquitlam and Vashon stades, the latter representing the maximum of the Fraser Glaciation. Ice advance units are separated by thin organic sediments deposited during the Port Moody Interstade (Lian and Hickin 1993, 1996; Hicock and Lian 1995; Hicock et al. 1999; Lian et al. 2001). In both Lynn and Seymour valleys the glacial sediment fill rests locally on a sequence up to 5 m thick consisting of gravel, organic-rich silt and sand, and compressed woody peat that was deposited immediately before the last glaciation. This nonglacial sequence is the focus of our study and rests locally on diamicton, which in turn rests on bedrock. Late Wisconsinan stratigraphy in the Port Moody region is similar to that observed in Seymour valley (e.g., Hicock 1976; Hicock and Armstrong 1981; Hicock and Lian 1995). Of interest to this study are organic-rich sediments that occur stratigraphically below Fraser Glaciation sediments behind Port Moody Secondary School (Fig. 1).

## Methods

Radiocarbon samples were collected and cleaned of obvious contaminating debris in the field and put immediately into sterile containers. The samples that were radiocarbon dated at the W.M. Keck Carbon Cycle Facility (those with UCIAMS lab numbers, Table 1) by accelerator mass spectrometry (AMS) consisted of small wood fragments (<50 mg). The wood fragments were extracted from small blocks of peat, cleaned and examined for contamination by microscope, at Paleotec Services, Ottawa, Canada before being submitted for dating. All of the other radiocarbon samples collected during this research were selected at the section and consisted of larger wood fragments that were dried and sent directly to either the Geological Survey of Canada Radiocarbon Laboratory or the Waikato Radiocarbon Dating Laboratory. These samples were dated using conventional methods, except for Wk-19195 which was dated by AMS. In all cases samples received standard acid-base-acid treatments before combustion.

Pollen samples were collected from cleaned exposures and immediately placed in sterile containers. Samples of approximately equal volume were prepared and counted, and data compiled and presented following standard methods (Moore et al.1991; Hebda 1995). These included 5% hot KOH treatment of organic samples, HF pre-treatment for all mineral samples and 5 minutes of acetolysis. Resulting residues were screened at 10 micrometres. Palynomorphs were identified using standard keys such as Moore et al. (1991) and reference to a comprehensive reference collection at the Royal British Columbia Museum, Victoria, BC. Identifications were made at 400x magnification and 1000x under oil immersion for critical determinations. Where possible 300 grains or more were counted including Cyperaceae in the sum. Data were compiled, percentages calculated and diagrams plotted using PSIMPOL software which included stratigraphically constrained cluster analysis (CONISS) (Bennett 2002).

## Results and Interpretation

### Study sections: lithostratigraphy and chronology

#### *Lynn Valley*



Pre-Fraser organic-rich deposits in Lynn Valley have been studied for more than half a century (Draycott 1948; Dyck and Fyles 1962; Lowdon et al. 1967; Lowdon and Blake 1979, 1981; Clague 1980; and McNeely and Atkinson 1995). As observed in our study, the sequence at the west side of Lynn Creek (LW) consists of about 1 m of compact peat containing one thin sand bed about 2 cm thick (Armstrong et al. 1985; Armstrong 1990: 87-88). The peat unit is underlain by a 10-cm thick silt unit, which in turn rests on diamicton. The peat is capped with about 60 cm of silt and sand, truncated by Fraser Glaciation sediments. Four finite radiocarbon ages indicate that the peat unit at the LW section was deposited from ~ 47.8 to 33.0 $^{14}$C ka (Fig. 2; Table 1) representing about 15 000 years of nearly continuous accumulation.

A correlative section on the east side of Lynn Creek (LE) is exposed in an active cut-bank. Divigalpitiya (1982) and Huntley et al. (1983), used mineral sediments from the peat to test thermoluminescence dating, and Lian et al. (1995) tested optical dating protocols on it. Stratigraphically, the LE and LW sequences are similar. At its base the LE sequence has almost 1.5 m of nearly massive matrix- and clast-supported diamicton that rests directly on bedrock. Diamicton stone shapes range from subrounded, with rare worn and faint striae, to subangular and lacking striae. Near the base of the diamicton unit (Fig. 2) sorted sediments are conformably interlayered with the diamicton; in other places they are sub-horizontally to crudely cross-bedded, including sand and laminated silt lenses. The diamicton is overlain by 50–80 cm of organic-rich silt and compressed woody peat (Fig. 3), including a conspicuous 5 cm-thick sand bed about 20 cm above the lower contact. Wood near the base of the peat unit at LE gave an age of 45.0 $^{14}$C ka BP, and wood near the top of the unit yielded an age of 28.0 $^{14}$C ka BP (Fig. 2; Table 1). Peat is overlain by about 40 cm of horizontally-bedded silty sand, locally interbedded with layered disseminated organics including wood fragments near the top. Wood from 4 cm and 20 cm below the top of the unit yielded ages of 26.7 and 27.1 $^{14}$C ka BP, respectively (Fig. 2).



***Seymour Valley***

The Seymour Valley (SV) section occupies an active cut bank on the west side of Seymour River (Fig. 1) (Lian and Hickin (1993). About 5 m thick, the section consists generally of three upward-fining sequences, each consisting of up to 1 m of moderately-sorted and moderately to well-rounded pebble gravel and sand at the base, overlain by horizontally bedded silt, sand, and fine gravel (Fig. 4). The silt and sand beds commonly contain disseminated organics. Compressed woody peat occurs between coarse units. Bulk peat from near the base of the sequence yielded an age of 41.1 $^{14}$C ka BP, and wood from compressed peat near the centre of the section produced an age of 37.1 $^{14}$C ka BP (Fig. 4; Table 1). Two wood samples from an organic-rich silt ~3 m higher in the section yielded ages of 35.1 and 35.7 $^{14}C$ ka BP. Bulk peat from the uppermost peat bed gave an age of 29.4 $^{14}C$ ka.

***Port Moody***

The Port Moody (PM) section is located behind the Port Moody Secondary School (Lowdon and Blake 1978, Hicock 1980). It consists of about 4 m fine to medium sand, which is overlain by 5 cm of silt then 30 cm to 1 m of compressed and faulted (up to 2 m displacement) fissile peat. A bulk peat sample yielded age of 31.0 $^{14}C$ ka BP (Table 1). Peat is conformably overlain by 5 cm of clayey silt, then ~3 m of medium sand and gravel.

**Study sections: depositional history**



The two Lynn Valley sequences, above the basal diamicton, are together interpreted to represent nearly continuous peat accumulation in a floodplain back swamp between 47.8 and 28.0 $^{14}C$ ka BP. A minor fluvial incursion into the back swamp wetland deposited the sand bed about 36.3 $^{14}C$ ka BP. After 28.0 $^{14}C$ ka BP peat accumulation was replaced briefly by fluvial sand and silt deposition which persisted until after 26.7 $^{14}C$ ka BP. The diamicton at the base of the LE section is either an immature till (cf. Lian and Hicock 2010) deposited during the Semiahmoo glaciation (MIS4), or perhaps a paraglacial alluvial deposit formed at the end of the Semiahmoo glaciation or at the beginning of the Olympia Interstade.

The Seymour Valley and Port Moody sequences were also deposited in a fluvial setting. Seymour Valley sediments are interpreted as a back-swamp succession representing about 5 m of floodplain aggradation over ~12 000 years. They were laid down at the same time as the Lynn valley sequences, but in a more fluvially proximal, active environment. At Port Moody wetland deposits were overlain and deformed by an ice contact complex then ~2 m of diamicton interpreted to be Coquitlam till. The ~15 m of sand, gravel and diamicton are interpreted as further ice-contact deposits which are then capped by ~2 m of diamicton (Vashon till).

**Palynology**

The most comprehensive description of pollen zones is for LW which exhibits the longest record (Fig. 5). Assemblage zones are described separately from their interpretation to facilitate comparison and correlation with records outside the Fraser Lowland, such as those on the Olympic Peninsula, and to establish a reference stratigraphy for the Olympia Interstade. Chronology of the upper pollen zones in LW section is inferred from more recent radiocarbon ages obtained from the LE section (Fig. 4). Vegetation and climate are interpreted on the basis of all four records, which represent more or less contemporaneous landscapes. This approach enables the recognition of local versus regional vegetation changes. Polypodiaceae spore values are expressed as a percent of total pollen and spores but are excluded from the sum.

***Lynn Valley west (LW)***

Six pollen zones are identified by visual inspection of the pollen spectrum for LW (Fig. 5) and are generally confirmed by CONISS zonation. Considering the stratigraphy and radiocarbon ages, the pollen assemblages appear to span the Olympia Interstade. Also represented are the non-arboreal assemblages found in sediments deposited immediately before and after it.

Pollen zone LW-1 (Polypodiaceae-NAP, >48 $^{14}$C ka BP) is represented by only two samples and is dominated by Polypodiaceae spores. *Lycopodium alpinum* type spores are also characteristic and grass and alder pollen are the other main components. Asteraceae and Cyperaceae pollen are notable whereas arboreal pollen is nearly absent. In zone LW-2 (*Pinus-Picea-Tsuga mertensiana,* >48 to ~44 $^{14}$C ka BP) AP dominates with *Pinus* then *Picea* and *T. mertensiana,* respectively. All identifiable *Pinus* pollen is of the *P. contorta* type. Poaceae pollen is the most abundant NAP, and the occurrence of *Valeriana* pollen is notable. *Lycopodium* and Polypodiaceae spores are almost absent compared to their abundance in zone LW-1.

In zone LW-3 (*Pinus-Tsuga heterophylla,* ~44 to ~39 $^{14}$C ka BP) AP continues to dominate with *Pinus* predominant, but *T. heterophylla* (10–20%) is also abundant. In zone LW-4 (*Picea-Tsuga mertensiana-Cyperaceae,* 39 to 27.5 $^{14}$C ka BP) *Picea* and *T. mertensiana* again co-dominate the assemblage but with a notable component of Cyperaceae (10–30%). In the first half of the zone *T. mertensiana* dominates along with notable amounts of *Myrica*, whereas *Picea* dominates in the second half.



Zone LW-5 (Poaceae-*Alnus*-Polypodiaceae 27.5 to <27 $^{14}$C ka BP) begins with a sharp rise in Poaceae pollen (>50%) and is accompanied by an increase in *Alnus (*10%). Ericales, Asteraceae, *Caltha* and *Gentiana* pollen occur in relative abundance. AP values drop to less than 10% at the top of the zone. *Lycopodium* and Polypodiaceae spores are exceptionally numerous compared to the rest of the record. The two NAP-dominated samples of zone LW-6 differ from those in zone LW-5 mainly because of increased Cyperaceae and much less Polypodiaceae.

***Lynn Valley east (LE)***

The LE sequence is divided into six assemblage zones (Fig. 6) which resemble those in the LW sequence, but lacks a basal NAP zone or a well-developed NAP zone at the top. In zone LE-1 (*Pinus*-Poaceae, >45 to ~36 $^{14}$C ka BP) *Pinus* dominates (40–50%) with a notable component of Poaceae, as well as *Alnus, Abies* and *Picea*. The highest *Tsuga heterophylla* values (>10%) for the sequence occur in this zone. The zone LE-2 assemblage (*Tsuga mertensiana*-*Picea*-Cyperaceae, ~36 to 27.5 $^{14}$C ka BP) has co-dominants *Tsuga mertensiana* and *Picea*, interrupted by a strong *Pinus* peak (45%) late in the zone. Cyperaceae (~10–40%) and Poaceae (~10–20%) occur abundantly with a notable admixture of *Gentiana* pollen (~1–5%).

In zone LE-3 (*Tsuga mertensiana*-*Picea*-Polypodiaceae-*Lycopodium*, 27.5-27.1 $^{14}$C ka BP) the AP component strengthens compared to that in zone LE-2 to more than 50% at the cost of Poaceae and Cyperaceae. Polypodiaceae spores occur abundantly (up to 50%) and *Lycopodium* values exceed ~10%.

In zone LE-4 (*Picea*-Poaceae-Cyperaceae-*T. mertensiana*: ~27.1 to 26.5 $^{14}$C ka BP) *Picea* pollen (20–30%) dominates slightly, and Cyperaceae (15–25%) and Poaceae (10–25%) pollen are abundant. *T. mertensiana* persists, decreasing from the preceding zone, whereas Polypodiaceae spores occur infrequently. In the two-sample zone LE-5 (Cyperaceae-Poaceae, <26.5 $^{14}$C ka BP) the first sample is dominated by Cyperaceae whereas the second sample is dominated by Poaceae. The relatively small AP pollen signal consists mostly of *Picea*. The single sample in zone LE-6 shows a sharp increase in Poaceae at the expense of all other types.

***Seymour Valley (SV)***

In this sequence of highly mixed lithology and irregularly spaced samples, only three zones are recognized (Fig. 7). Zone boundary dates assume that gravel units were deposited in relatively short intervals compared to silts and organic beds during which sedimentation was assumed to be relatively constant.

In zone SV-1 *Pinus* and Poaceae pollen dominate the spectra (*Pinus*-Poaceae- *T. mertensiana*: 41 to ca. 37 $^{14}$C ka BP), but *Pinus* values reach a maximum of only ~ 35%. *T. mertensiana* occurs up to 30%. Notable also are Polypodiaceae (up to 20%) and *Cryptogramma* (up to 5%) spores. The basal sample is dominated by NAP - mainly Cyperaceae, Poaceae and Lamiaceae. In zone SV-2 (*T. mertensiana-Picea*: ca. 37 to ca. 36 $^{14}$C ka BP) *T. mertensiana* and *Picea* pollen dominate, but the NAP component is abundant and diverse. Included are *Salix*, Asteraceae, and *Sanguisorba,* all at least 5%. In zone SV-3 (*Picea*-Cyperaceae: ca.36 to <29 $^{14}$C ka BP) AP values are in the 20–30% range with *Picea* dominating. Cyperaceae values vary widely and reach 30–50%. The wetland shrub *Myrica* is notable. Poaceae are lower in abundance than in the previous two zones.

***Port Moody (PM)***

The sparse basal sample of zone PM-1 (*Alnus*-Poaceae: age unknown) is dominated (>50%) by *Alnus crispa* type pollen with Poaceae ~15% (Fig. 8). Conifer pollen is almost absent in the mid portion (zone PM-2) (>31 $^{14}$C ka BP) and Cyperaceae (10–35%) pollen dominates, with *Picea* and Poaceae (5–20%) as secondary types. A diversity of infrequent but consistent herbaceous meadow types is noteworthy including: *Artemisia*, Apiaceae*, Sanguisorba* and *Polygonum*. The upper portion of the sequence (zone PM-3, ~31 $^{14}$C ka BP and younger) has *Picea* pollen reaching >40% with *T. mertensiana* as a secondary conifer. The NAP portion is dominated by grasses and fern spores and includes numerous Asteraceae and *Sanguisorba* in the top-most sample.

**Vegetation, Landscape and Climate**

The reconstruction of environments and events is based primarily on the two continuous and well-dated records in the LE and LW sections with variations from the SV and PM sections noted. The chronology is based on the LW section in the lower part and on the LE section in the upper portion where there is better chronologic control and more resolution in pollen zones (Fig. 2). The correlations are: zones LW-1 and LW-2 have no equivalents in LE; zones LW-3 and LE-1 are more or less equivalent; zone LW-4 encompasses zones LE-2, -3, and -4; zones LW-5, and -6 include zone LE-5 and extend beyond it in time.

The non-arboreal pollen assemblage at the base of the LW sequence (zone LW-1) reveals a tundra or tundra-steppe landscape before 48 $^{14}$C ka BP. Climate was certainly cold, but whether it was moist or dry is uncertain in part because of the abundance of Cyperaceae pollen - an indicator of edaphically moist sites.

High AP pollen values starting before 48 $^{14}$C ka BP signal warming and forest development (zone LW-2; possibly early zone LE-1). The diversity and abundance of conifer pollen suggests closed to partly open mixed conifer forests perhaps resembling those that occurred widely in late glacial times on the west coast of North America (Hebda and Whitlock 1997; Walker and Pellatt 2008). The pollen assemblage, particularly the abundance of *T. mertensiana,* suggests a cool climate (Pellatt and Mathewes 1997), not as warm as present but much warmer than that inferred for the preceding interval. Relatively high pine pollen values and dominance of grasses rather than Cyperaceae in the NAP, imply only moderate moisture availability. Abundant Cyperaceae pollen is usually an indicator of local wetlands that can occur even when upland conditions are comparatively dry. It is not clear whether grasses grew at the site of deposition as dominant wetland plants or in adjacent upland openings. Moist open to partly shaded patches are suggested by *Valeriana* pollen.

The warmest climate in the Olympia, as recorded at our sites, is indicated by relatively abundant *Tsuga heterophylla* pollen in zone LW-3 (and in zone LE-1), presumably reflecting growth of the species in the forest stands (see surface sample data in Hebda and Allen 1993, Allen et al. 1999) during the ~44 to 39 $^{14}$C ka BP interval. Considering its mixing with pollen of *T. mertensiana* and other conifers, a climate like that at the mid-elevation transition between today's Coastal Western Hemlock and Mountain Hemlock (MH) biogeoclimatic zones likely prevailed. Accordingly, ecological zones were depressed 1200 to 1500 m compared to today (Meidinger and Pojar 1991), much more than suggested for the same time in the highly oceanic Haida Gwaii (Warner et al. 1984).

*Picea* -*T .mertensiana* forests or parkland occurred from 39 to 26.7 $^{14}$C ka BP, dotted with sedge- and *Myrica* -dominated fens. The abundance of the pollen of these two wetland indicators suggests a moist climate, and the relatively low abundance of *T. heterophylla* indicates cool conditions presumably similar to the inland variants of the MH zone today. The *Picea* pollen is inferred to belong to *Picea engelmannii* not *Picea sitchensis* (Sitka spruce) because the study area was likely well inland from the ocean (Clague 1976), antecedent and subsequent climate was dry and later macrofossils of *Abies lasiocarpa* of Port Moody Interstade age from nearby sites indicate continental climates (Hicock et al. 1982). However, the possibility of *P. sitchensis* as the source of the pollen cannot be ruled out.

A sharp decline in AP values reveals that forest cover appears to end abruptly at the start of zones LW-5 and LE-5, just after 26.7 $^{14}$C ka BP, based on new radiocarbon ages obtained from the LE section (Fig. 2; Table 1), as spruce-mountain hemlock forest was replaced by grassy steppe, perhaps with widely scattered clusters of trees. The diversity of pollen and spore types suggests a rich herbaceous flora including species in the Asteraceae (aster) and Apiaceae (parsley) families. Ferns may have dominated in moist sites. Overall, though, climate became cold and dry. This abrupt change occurred much later than interpreted from the previous dating of the LW exposure (26.7 compared to 33 $^{14}C$ ka BP in Armstrong et al. 1985), the difference is presumably due to erosion at the top of the peat unit in exposure LW.

The floodplain near the location of the Seymour Valley section was much more active than that near the Lynn Valley sections as reflected by frequent accumulation of clastic sediments of widely varying texture in the former, which allowed only thin peat layers to develop. Before 37 $^{14}C$ ka BP *Pinus* – *T. mertensiana* stands covered the area with less *Picea* than in nearby Lynn Valley. Considering the abundant Poaceae pollen in the assemblages, relatively dry openings were widespread, perhaps reflecting a more open forest than in Lynn Valley. The low values for *T. heterophylla* are notable, suggesting three possible explanations: (i) the interval of its abundance is missing at Seymour Valley, (ii) local climate was drier or colder than in Lynn Valley because of cold air drainage in the much longer and larger Seymour Valley, or (iii) the dominantly coarse-textured surficial sediments perhaps associated with an anastomosing river system resulted in edaphic conditions unsuitable for the species.

After 37 $^{14}C$ ka BP the landscape in lower Seymour Valley was covered in *Picea* and *T. mertensiana* stands with widespread fen openings similar to adjacent Lynn Valley. *Myrica* was also abundant in these wet openings. The inferred cool moist climate is consistent with that deduced for Lynn Valley despite the highly varying sedimentary regime.

Except for the basal sample, the deposits at Port Moody record an open *Picea* forest with widespread Cyperaceae-dominated openings (presumably fens) similar to conditions at the other two sites. The basal sample at Port Moody, dominated by alder with NAP, is unlike any other encountered in this study. The assemblage may reflect an unstable (*Alnus*) landscape with numerous meadow openings under a cool to cold climate. The contact of the sediments enclosing the sample with overlying fine-grained deposits is somewhat abrupt suggesting a depositional hiatus. Accordingly, the basal assemblage at Port Moody is interpreted to represent an earlier time than represented by the continuous sequence at Lynn Valley, a time presumably before 50 $^{14}C$ ka BP.

## Discussion

Our Fraser Lowland sequences provide an opportunity to compare the nature and timing of events with other terrestrial sites in the region, and with off-shore marine sequences, which can help us understand changes in the broad pattern of vegetation and climate. In particular our records allow a look at the similarities and differences between CIS proximal (inland) and distal sites during a complete glacial - nonglacial - glacial cycle (Fig. 9). These comparisons, placed in the context of environmental changes outside the region, help discern whether or not the events in the Fraser Lowland were driven by local factors such as proximity to mountain ice, or by much broader hemispheric changes as reflected in wide-ranging vegetation variation and changes in climatic proxies such as oxygen isotope ratios (i.e. Stuiver and Grootes 2000; Jimenez-Moreno et al. 2010). These analyses help clarify the climatic conditions of the Olympia Interstade and define, on land, its bounding ages in comparison to marine and ice core sequences. They also establish a palynostratigraphic framework for future comparisons and dating.

**Comparison to pollen records beyond the CIS limit**

Our sequence exhibits similarities to, and major differences from, sequences in the region (Figs. 1, 9). Comparisons are made with the proviso that the dating of vegetation boundaries is poorly constrained in the early Olympia interval. Zone boundary ages in the Humptulips (on Washington State's Olympic Peninsula) sequence are difficult to place because of low sampling resolution (5 cm) during slow sediment accumulation (Heusser et al. 1999; Gavin and Brubaker 2015).

On the Olympic Peninsula there is a >38.5 $^{14}$C ka BP grass-NAP dominated zone extending beyond the limit of radiocarbon dating (Gavin and Brubaker 2015: Fig 4.1 zone H1C-7) that may represent the pre-Olympia stadial and may also include the relatively dry and cool early part of the Olympia (equivalent to most of zone LE-1) (Fig. 6). At about the same time at Carp Lake in the eastern Cascade Mountains of Washington State, boreal and temperate conifers were mixed with *Artemisia* openings reflecting a cooler and more humid climate than today (Jimenez-Moreno et al. 2010). At Fargher Lake, southern Washington State pine-fir-mountain hemlock parkland beyond the range of radiocarbon dating, is replaced by tundra-like vegetation dominated by grasses 50-43 $^{14}$C ka BP and followed by mixed conifer forest (Grigg and Whitlock 2002).

Generally there was a geographically extensive cool to cold episode before the onset of the relatively warm conditions of the Olympia Interstade (Olympia optimum). When the pre-Olympia cold interval ended is hard to identify because of lack of reliable dating. An interval of relatively warmer pine–grass landscape persisted in the Fraser Lowland for several thousand years beginning before 48 $^{14}$C ka BP and ending by about 44 $^{14}$C ka BP before the Olympia optimum ( *T. heterophylla* pollen maximum). Grasses were relatively prominent and the climate was drier and cooler than today.

Following this interval, the *T. heterophylla* pollen peak in the Fraser Lowland indicates the warmest part of the Olympia Interstadial. A strong *Tsuga heterophylla* signal starting at about 38 $^{14}C$ ka BP on outer coast of the Olympic Peninsula (Heusser 1977, Heusser et al. 1999) seems to mark the warmest point there. At Little Lake in Oregon the warmest interval, including peaks of *T. heterophylla* and *Pseudotsuga,* appears to occur slightly later (Grigg et al. 2001). At Fargher Lake, the interval of relatively abundant *Pinus* and Poaceae of zone FL1c (ca. 49-41 $^{14}C$ ka BP) precedes an interval of varying but marked *T. heterophylla* peaks (zone FL-2) (Grigg and Whitlock 2002). A strong Cupressaceae signal is associated with the beginning of *T. heterophylla* increases and *Abies* occurs abundantly in the zone. Zone FL1c is interpreted as cold and dry parkland and considering the high grass and NAP, might even qualify as tundra or cold steppe. In zone FL-2 the relatively high frequency variation of *T. heterophylla,* versus *Pinus* and Poaceae, is interpreted as shifts between high and mid elevation forests. This vegetation, and presumed climatic variation, persisted until 31 to 32 $^{14}C$ ka BP. Zone FL-1c would seem to match most of LW-2, and zone FL -2 appears to correlate with zones LW-3 and LW-4 (Fig. 5).

Based on these correlations the warmest interval in the Fraser Lowland during the Olympia Interstade appears to have been shorter than at Fargher Lake. In the latter half of the forested interval, *Picea* and *T. mertensiana* dominated in the Fraser Lowland (early part of zones LW-4 and LE-2) as might be expected today at more northerly latitudes and possibly closer to high elevation ice masses. At the same time moist conifer (montane) forests occurred in the Coast Range of Oregon (Worona and Whitlock 1995; Grigg et al. 2001) and open forest of boreal and temperate species grew under cool and dry conditions east of the Cascades (Carp Lake; Whitlock and Bartlein 1997; Jimenez-Moreno et al. 2010).

From ~39 to 26.7 $^{14}C$ ka BP cool, moist climate prevailed in the Fraser Lowland unlike several sites south of the CIS (Heusser 1977; Heusser et al. 1999; Grigg and Whitlock 2002). Olympic Peninsula sites exhibited a strong NAP signal, especially grasses, some sedges and pine

(Humptulips sequence), and spruce. The Fargher Lake record had high frequency variation on a millennial scale of AP and NAP types (grasses mainly) (Grigg and Whitlock 2002) as did to some extent Little Lake (Grigg et al. 2001). These varied assemblages at different sites reveal strongly differing climates over relatively short distances. These differences may be the result of relatively low resolution sampling at Fraser Lowland sites.

Cold and dry climates arrived after 26.7 $^{14}$C ka BP in the Fraser Lowland and forest and parkland were replaced by open tundra-like vegetation. Cascade Mountain sites (Carp and Fargher lakes) indicate onset of cold dry climates with the development of grassland steppe or tundra at about the same time (see zone FL-3b in Grigg and Whitlock 2002) with a lead-up interval of cooling of about 4–5 ka (see zone FL3a in Grigg and Whitlock 2002). A short lead up cooling is evident in our Fraser Lowland zones LE-3 and -4 during which rising grass and Cyperaceae values signal decreasing tree cover. In contrast, coastal sites remain largely forested (*T. mertensiana*) and cool, but not as cold and dry as at the inland sites. At Washington State coastal sites, cold and dry conditions begin developing at about 31 $^{14}$C ka BP (Fig. 9).

There are clear differences in pollen assemblages and interpreted vegetation during the middle to latter Olympia Interstade between the Fraser Lowland and coastal Olympic Peninsula even considering Gavin and Brubaker’s (2015:65) suggested alternate time scale. The differences to some degree may be the result of the relatively low temporal resolution of Fraser Lowland sequences. But they may also be related to a strong coastal-interior climate gradient. Another possibility is that inland ecosystems were simply not sensitive enough to respond to the high frequency cooling and warming in the interval, whereas the coastal ecosystems were. Alternatively, open vegetation on Olympic Peninsula may have resulted from relative drought rather than cold; the abundance of grasses throughout the Olympia interstadial in the Humptulips record is notable in this respect (Gavin and Brubaker 2015: Fig.4.1).

**Comparison to pollen and marine sediment records north of the CIS limit**

Within the limits of the CIS, pollen assemblages on the east side of Vancouver are dominated by *Picea* before ~33 $^{14}$C ka BP and then *Alnus* to about 29 to 30 $^{14}$C ka BP after which NAP becomes more abundant (Alley 1979). Alley (1979) considered the pre-29 $^{14}$C ka BP climate to be similar to present based on comparison to modern pollen spectra and occurrence of *Pseudotsuga* pollen and wood. Our sites clearly do not reflect such conditions. The dry east Vancouver Island region might have supported *Pseudotsuga.* The relatively abundant grasses at Lynn Valley suggest that our area was also dry enough for *Pseudotsuga* yet it did not occur. Notably, *Pseudotsuga* can grow in much cooler-than-present conditions outside the region (Meidinger and Pojar 1991).



Sites on the more northerly Haida Gwaii (Warner et al. 1984) begin with a cold non-forested assemblage before 46 $^{14}$C ka BP, after which moist cool climate characterized by *Picea* –*Tsuga* forest with moist openings prevailed. Limited cooling and perhaps drying occurred after this interval with an increase in grasses but *Picea* and *Tsuga* remained. To the east Olympia-age pollen and spore sequences indicate *Picea*- or *Pinus*- dominated forest or woodland (Alley et al. 1986, Clague et al. 1990) in south-central BC possibly contemporaneous with tundra or cold-steppe communities to the northwest in central BC (Plouffe and Jetté 1997). Though the two most complete Olympia records from this region have complex and possibly interrupted sedimentary sequences (Alley et al. 1986: Figs. 2, 4; Clague et al. 1990: Figs 6, 10) it appears that at no time was there a tundra-like zone that reflects return to cold climate in the middle of the Olympia Interstade. In fact these sites have pollen assemblages with abundant spruce and sedges, remarkably similar to our Fraser Lowland sequences of the same age.

A recent, and well-dated, high resolution sequence of marine sediment collected ~70 km off the west coast of Vancouver Island (core MD02-2496, Fig. 1) provides a valuable comparison with respect to dating and the general pattern of Olympia events in the region (Chang et al. 2008, Cosma et al. 2008, Cosma and Hendy 2008). In Table 1 we have calibrated our radiocarbon ages using Oxcal 4.2 and the IntCal 13 data set; the ages given in the text are the median values (see also Fig. 9 for correspondence of $^{14}$C ages to cal ka BP ages). Beginning likely before 45 $^{14}$C ka BP, the record reveals a glacial/non-glacial/glacial sequence apparently spanning the full interval described in our study. Glacial climates are inferred to have occurred before 49 $^{14}$C ka BP ending much later than inferred from our terrestrial records in the Fraser Lowland. Generally, glacial episodes are associated with glaciomarine sediments and ice rafted debris (IRD). The non-glacial episode is associated with hemipelagic sediment. According to Cosma et al. (2008: 951) the shift to the non-glacial interval (beginning of Olympia) occurred between 41.1-38.4 $^{14}$C ka BP (Table 1). Even considering the issues with dating, this change-over seems to have occurred much later than the >48 $^{14}$C ka BP suggested by our terrestrial record. One possibility is that in general the climate was cool enough to maintain major glaciers in today's Strait of Georgia area and hence the glacial influence detected at the site of core MD02-2496, while forest or parkland was widespread on uplands. Pollen Zone LE-1 >45 to ~39 $^{14}$C ka BP reflects cool and possibly dry climate and fits the timing well. Cosma et al.'s (2008) hemipelagic marine interval from 38.4- 26.5 $^{14}$C ka BP also fits well with our interpretation of the warm part of the Olympia Interstadial, and is more or less coincident with spruce-hemlock forested pollen zones (LE2-4, LW3-4). Glaciomarine sedimentation began again about 26.5 $^{14}$C ka BP, remarkably close to the 26.7 $^{14}$C ka BP time for the onset of non-arboreal vegetation and cold climate in the Fraser Lowland.

Our Olympia sequences do not resolve the short duration "interstadials" observed in the MD02-2496 core (Chang et al. 2008, Cosma et al. 2008), in part because the resolution at our sites is much coarser and in part because the terrestrial environment may not have been as sensitive to rapid and short-lived marine cooling or valley ice-advances. The warmest interval (based on $\delta^{18}$O) recorded by Chang et al. (2008) more or less corresponds with our forested zone LW-3.

**Timing and character of the Olympia Interstadial**



Taken together the terrestrial and marine records allow us to constrain and summarize the characteristics of the Olympia Interstade (Fig. 9). On land relatively warm conditions begin to develop from a cold episode sometime before 50 $^{14}$C ka BP. The precise timing of this is not clear because of the resolution of dating and because the vegetation response to warming may not have begun in all parts of the region at the same time. During an interval of more than 6,000 years climate remained relatively cool to cold but trees (especially pine and fir) were widespread mostly in parkland communities. Despite warming on land, glaciomarine conditions persisted in coastal waters and the occurrence of IRD suggests that ice reached tidewater during the early part of the interval. About 44 $^{14}$C ka BP or shortly thereafter, the marked increase in western hemlock at several of the sites signals the start of warmest portion of the Olympia and the onset of about 14,000 or more years of cool and moist climate. Despite Alley's (1979) interpretation, the climate was likely never as warm as today. Considering the Fraser Lowland record, the warmest climate in this interval occurred at the beginning. Glaciomarine sedimentation ceased during this time and tidewater glaciers were presumably absent in the region. Progressive cooling began after 30 $^{14}$C ka BP however trees remained widespread on the landscape. Not until about 26.5 $^{14}$C ka BP did cold and dry glacial conditions prevail leading to the widespread expansion of tundra/cold steppe, indicated by grasses. Glaciomarine sedimentation returned and the region entered the full glacial state of the Fraser Glaciation.

The staged entry into, and progressive exit from the warmest part of the Olympia Interstadial, raise questions about the definition of the Olympia Interstade's duration. The narrowest definition would include only the interval during which there is no evidence for ice or cold climates either on land or in water. According to these criteria the Olympia Interstade spans about 44 $^{14}$C ka BP to the marked appearance of cold climates at 26.5 $^{14}$C ka BP. This definition contradicts the notion that the start of the Fraser Glaciation began about 30 $^{14}$C ka BP with the onset of progressive cooling (Clague and James 2002). If we consider only the terrestrial record then the Olympia Interstade could be defined as beginning with the end of cold climate (as defined by NAP predominance) >50$^{14}$C ka BP and ending at the previously defined start of the Fraser Glaciation at about 30 $^{14}$C ka BP or alternatively with the reappearance of NAP dominated plant communities just after 26.7 $^{14}$C ka BP when cold conditions returned. For our sites the climate was certainly not glacial from >50$^{14}$C ka BP to 26.7 $^{14}$C ka BP, although the marine record indicates ice must have occupied valleys somewhere in the early part of this interval (Fig. 9). Despite the current notion (onset of cooling at 30 $^{14}$C ka BP) the marine record suggests there is no evidence of glacial activity (glaciomarine deposits) influencing coastal waters (Chang et al. 2008) at that time.

Considering the uncertainty in timing of the preceding cold (stadial) episode and variation in character and timing of the unambiguously non-glacial interval (Jimenez-Moreno et a1. 2010) a broader rather than narrower definition may be appropriate. Accordingly we suggest that the Olympia Interstade began with marked warming before 50 $^{14}$C ka BP and ended about 26.5 $^{14}$C ka BP with the sharp and clear onset of cold climate.

Jimenez-Moreno et al. (2010) addressed the question of millennial scale climatic variation and events through a synthesis of paleoecologic records in North America and in so doing identified some key horizons of change (climatic boundaries) at the start and end of the Olympia Interstade. According to them, key boundaries occurred at ~53 $^{14}$C ka BP (about 58 cal ka BP), from ~37 to 36 $^{14}$C ka BP (42 to 41 cal ka BP), and from ~27 to 28 $^{14}$C ka BP (31 to 32 cal ka BP). These boundaries are remarkably similar, considering the resolution of dating, to those for changes related to the Olympia Interstade, a pattern suggesting hemispheric control on major vegetation shifts even within the CIS limit.



On a hemispheric scale, persistent intervals of GISP2 $\delta^{18}$O values match very well with the inferred start and end dates for the Olympia Interstade (Stuiver and Grootes 2000) (Fig. 9). The marked and sudden rise in $\delta^{18}$O at about 53$^{14}$C ka BP (58 cal ka BP) correlates well with our estimate for onset of non-glacial conditions. This $\delta^{18}$O shift ushers in an interval of fluctuating but generally higher $\delta^{18}$O values. These persist with notable variations in the latter part of the interval until about 27-28 $^{14}$C ka BP (about 32 cal ka BP) (Fig. 9) when Fraser Lowland plant communities shifted from forested to non-forested biomes. Notably, similar biome shifts occurred at about the same times in southern Europe (Allen and Huntley 2000).

## Conclusions

Fraser Lowland pollen sequences reveal marked biome shifts from non-arboreal to coniferous arboreal vegetation at the start of the Olympia interval and arboreal to non-arboreal vegetation at the end of it. Broadly speaking, these changes correlate with vegetation shifts beyond the Cordilleran Ice Sheet limits in the region. However, Fraser Lowland sequences do not exhibit the high frequency vegetation changes interpreted in the later part of the Olympia south of the CIS. Nor do the Fraser Lowland records include non-arboreal vegetation in mid Olympia times as coastal Washington State sites do. On the basis of our analyses and hemispheric comparisons the Olympia interval was not an interglacial (as suggested by Alley 1979) - it was a long interstadial. Temperatures and associated vegetation never approached those of the Holocene and vegetation zones were depressed by hundreds of metres compared to present. Conditions similar to those of the late Pleistocene were reached in central west coast North America and even then for much of the Olympia they were not stable. The end of the Olympia was gradual, unlike its beginning, extending for several millennia until a full-glacial climate took hold about 26 to 28 $^{14}$C ka BP. We concur with Whitlock and Bartlein (1997) that, even within the limits of the CIS, interstadial vegetation was strongly shaped by hemispheric climatic variations. Further higher resolution and better dated sites within the CIS limits are needed to understand the differences in timing and types of changes south of the ice limit. Specifically, an explanation is needed for Olympia-aged contemporaneous open plant communities south of the ice sheet in coastal Washington State.



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Table 1

## Figures

Figure 1. Location of features and sites referred to in the text. Main map: LV = Lynn Valley sections; SV = Seymour Valley section; PMDS = Port Moody disposal pit section; PM = Port Moody Secondary School section. Inset map: BP = Bullion Pit; D = Dashwood; MC = Meadow Creek; PM = Pilot Mill ; CC = Cascade Cave; H = Humptulips; MC = marine core MD02-2496. Fargher Lake and Carp Lake are located ~50 km south of the map boundary, as indicated. Modified from Fig. 1 of Lian et al. (2001).

Figure 2. Lithostratigraphy and radiocarbon ages at the Lynn Valley sections. See Table 1 for a discussion of how the uncertainties associated with the GSC ages are reported. The location of radiocarbon sample GSC-93 is shown as a range because its precise position is unknown.

Figure 3. Lynn Valley east exposure showing the compressed peat bed interrupted by a sand bed about 5 cm thick, as it appeared in 2006. The peat bed is separated from the underlying diamicton by a unit of organic-rich colluvium likely derived from the diamicton. The peat is overlain by silty sand. The units on the scale bare are 1 cm.



Figure 4. Lithostratigraphy and radiocarbon ages at the Seymour Valley section. See Table 1 for a discussion on how the uncertainties associated with the GSC ages are reported.

Figure 5. Pollen and spore diagram, Lynn Valley west (LW)

Figure 6. Pollen and spore diagram, Lynn Valley east (LE)

Figure 7. Pollen and spore diagram, Seymour Valley (SV)

Figure 8 Pollen and spore diagram, Port Moody Secondary School (PM)

Figure 9. Comparison of selected pollen zones of the North American Pacific northwest region, the marine sequence from core MD02-2496 off the west coast of Vancouver Island, and the ice core sequence from GISP2, Greenland. GISP2 $\delta^{18}O$ modified and smoothed from Stuiver and Grootes (2000).

**Table 1.** Radiocarbon ages from study sites in Lynn valley, Seymour valley, and at Port Moody.

| Site | Lab number[a] | Material | $^{14}$C age, years BP[j] | Calibrated years BP[k] | Reference |
|---|---|---|---|---|---|
| Lynn, east | Wk-18971[b] | Wood | 26 730 ± 356 | 30 885 (30 240–31 377) | This paper |
| Lynn, east | Wk-20076[b] | Wood | 27 106 ± 266 | 31 111 (30 784–31 444) | This paper |
| Lynn, east | UCIAMS-54974[c] | Wood | 27 950 ± 120 | 31 644 (31 323–32 109) | This paper |
| Lynn, east | UCIAMS-54975[c] | Wood | 27 750 ± 180 | 31 485 (31 155–31 944) | This paper |
| Lynn, east | UCIAMS-61742[c] | Wood | 32 400 ± 210 | 36 297 (36 838–35 783) | This paper |
| Lynn, east | UCIAMS-75156[c] | Wood | 33 010 ± 260 | 37 141 (36 371–38 096) | This paper |
| Lynn, east | UCIAMS-75155[c] | Wood | 36 270 ± 400 | 40 909 (40 080–41 685) | This paper |
| Lynn, east | UCIAMS-61741[c] | Wood | 36 320 ± 330 | 40 966 (40 246–41 620) | This paper |
| Lynn, east | GSC-6843[b] | Wood (*Pinus*)[e] | 37 400 ± 600 | 41 836 (40 869–42 732) | This paper |
| Lynn, east | Wk-19195[c] | Wood | 44 956 ± 2200 | out of range | This paper |
| Lynn, west | GSC-2793[b] | Peat | 33 000 ± 310 | 37 152 (36 320–38 192) | Lowdon and Blake (1981) |
| Lynn, west | GSC-2873[b] | Wood (*Abies*)[f] | 34 900 ± 405 | 39 443 (38 596–40 367) | Lowdon and Blake (1981) |
| Lynn, west | GSC-93[b] | Wood | 36 200 ± 250 | 40 858 (40 261–41 416) | Dyck et al. (1965) |
| Lynn, west | GSC-3290HP[d] | Wood (*Tsuga*)[g] | 47 800 ± 550 | out of range | Lowdon and Blake (1981) |
| Port Moody | GSC-2533[b] | Peat | 31 000 ± 260 | 34 915 (34 416–35514) | Lowdon and Blake (1978) |
| Seymour | Beta-46053[b] | Peat | 29 440 ± 300 | 33 624 (32 945–34 149) | Lian and Hickin (1993) |
| Seymour | GSC-6879[b] | Wood | 35 100 ± 465 | 39 760 (38 670–40 752) | This paper |
| Seymour | GSC-5069HP[d] | Wood (*Picea*)[h] | 35 700 ± 160 | 40 315 (39 860–40 810) | Lian and Hickin (1993) |
| Seymour | GSC-5121HP[d] | Wood (*Abies*)[i] | 37 100 ± 170 | 41 651 (41 314–41 975) | Lian and Hickin (1993) |
| Seymour | GSC-6000HP[d] | Peat | 41 100 ± 300 | 44 622 (44 006–45 237) | This paper |

[a] Wk, Waikato Radiocarbon Dating Laboratory; UCIAMS, W.M. Keck Carbon Cycle Accelerator Mass Spectrometer Facility; GSC, Geological Survey of Canada Radiocarbon Dating Laboratory

[b] Conventional radiocarbon age
[c] Accelerator mass spectrometry (AMS) radiocarbon age
[d] Conventional radiocarbon age determined under high pressure, which extended the upper age limit achieved at the GSC Radiocarbon Dating Laboratory from approximately 40 000 $^{14}$C years BP to about 54 000 $^{14}$C years BP (Lowdon 1985).

[e] GSC Wood Report No. 2005-42; [f] GSC wood Report No. 79-28; [g] GSC wood Report No. 81-22; [h] GSC Wood Report No. 90-41;
[i] GSC Wood Report No. 90-74

[j] The Geological Survey of Canada (GSC) Radiocarbon Dating Laboratory reported ages with ±2σ uncertainties ("error" terms) while others are reported at ±1σ; the uncertainties associated with the GSC ages have been changed here to ±1σ for easier comparison with the other ages.

[k] Calibration was done using OxCal 4.2 (Brock-Ramsey 2009) and the IntCal 13 data set (Reimer et al. 2013). Median ages are shown, and the 2σ range is given in parentheses.

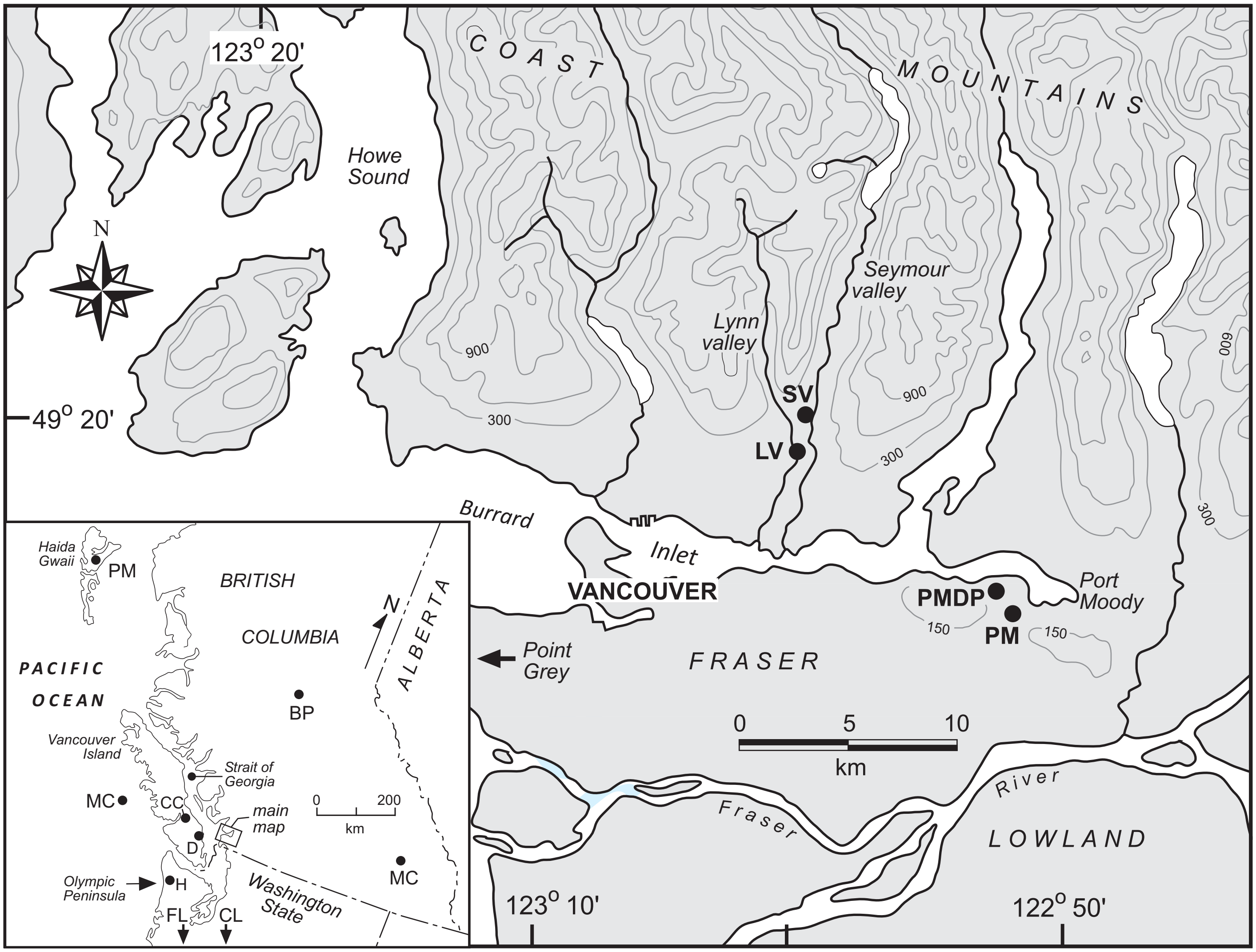
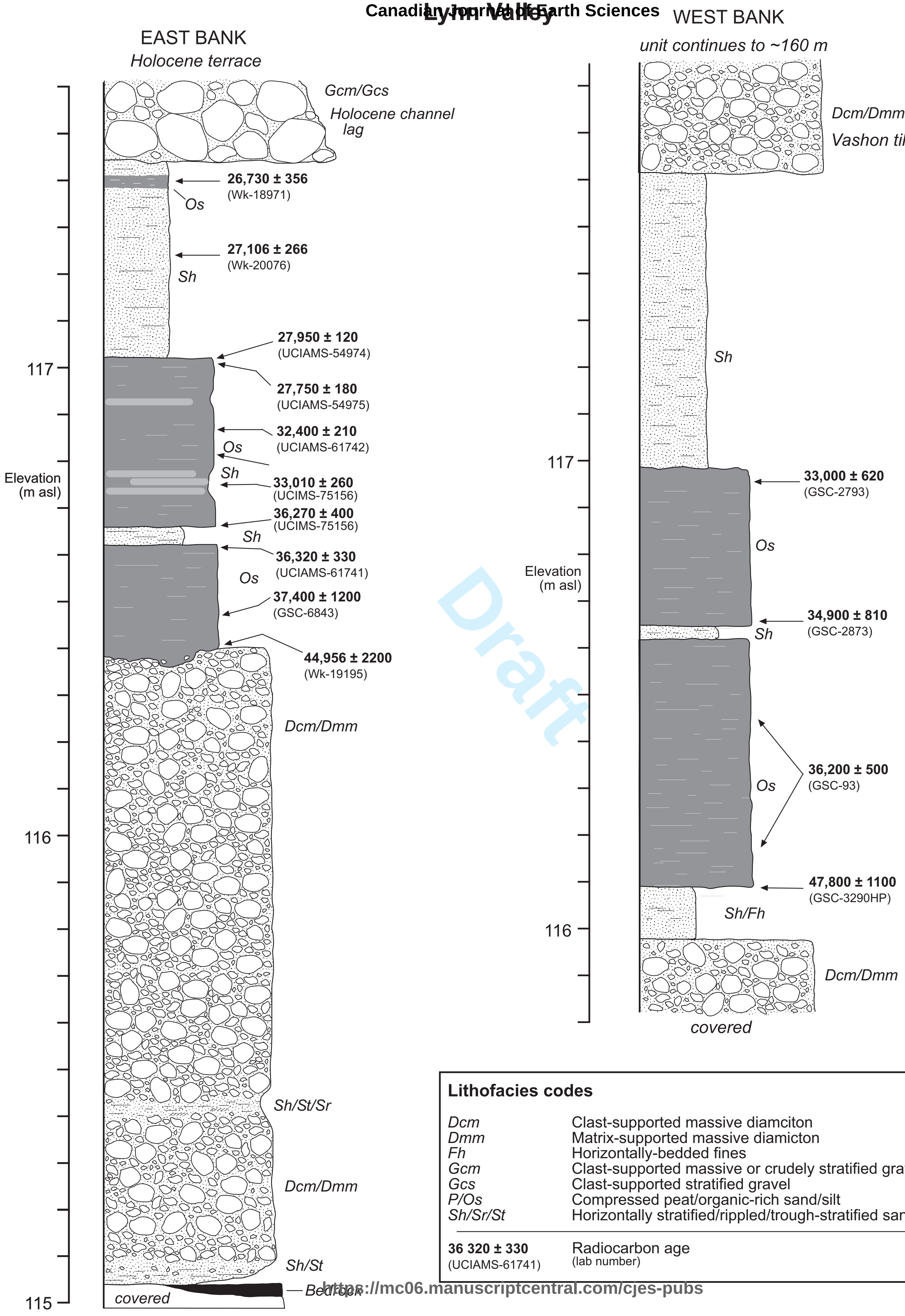



Fig. 1. (Hebda, Lian, and Hicock)




**Lithofacies codes**

| | |
|---|---|
| *Dcm* | Clast-supported massive diamciton |
| *Dmm* | Matrix-supported massive diamicton |
| *Fh* | Horizontally-bedded fines |
| *Gcm* | Clast-supported massive or crudely stratified gravel |
| *Gcs* | Clast-supported stratified gravel |
| *P/Os* | Compressed peat/organic-rich sand/silt |
| *Sh/Sr/St* | Horizontally stratified/rippled/trough-stratified sand |
| **36 320 ± 330** (UCIAMS-61741) | Radiocarbon age (lab number) |

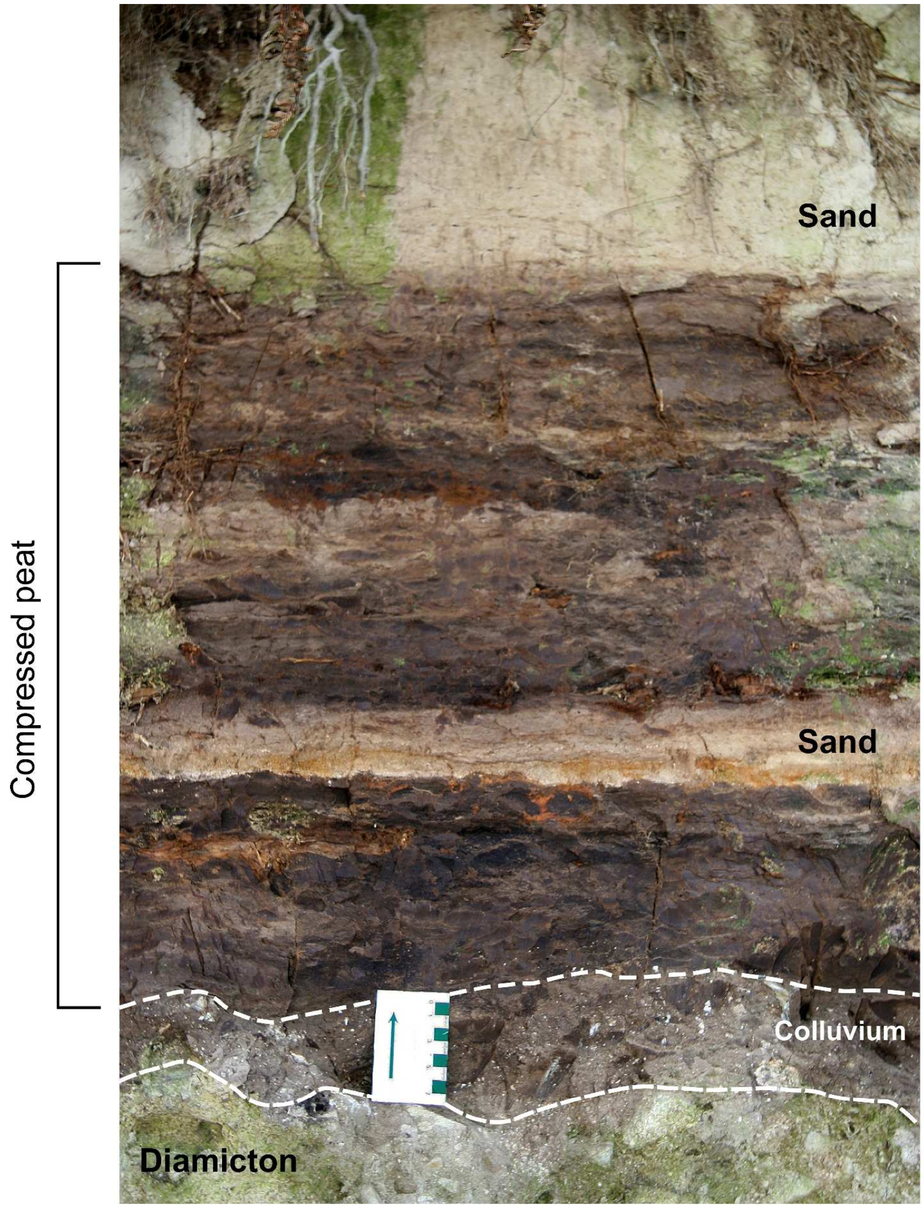
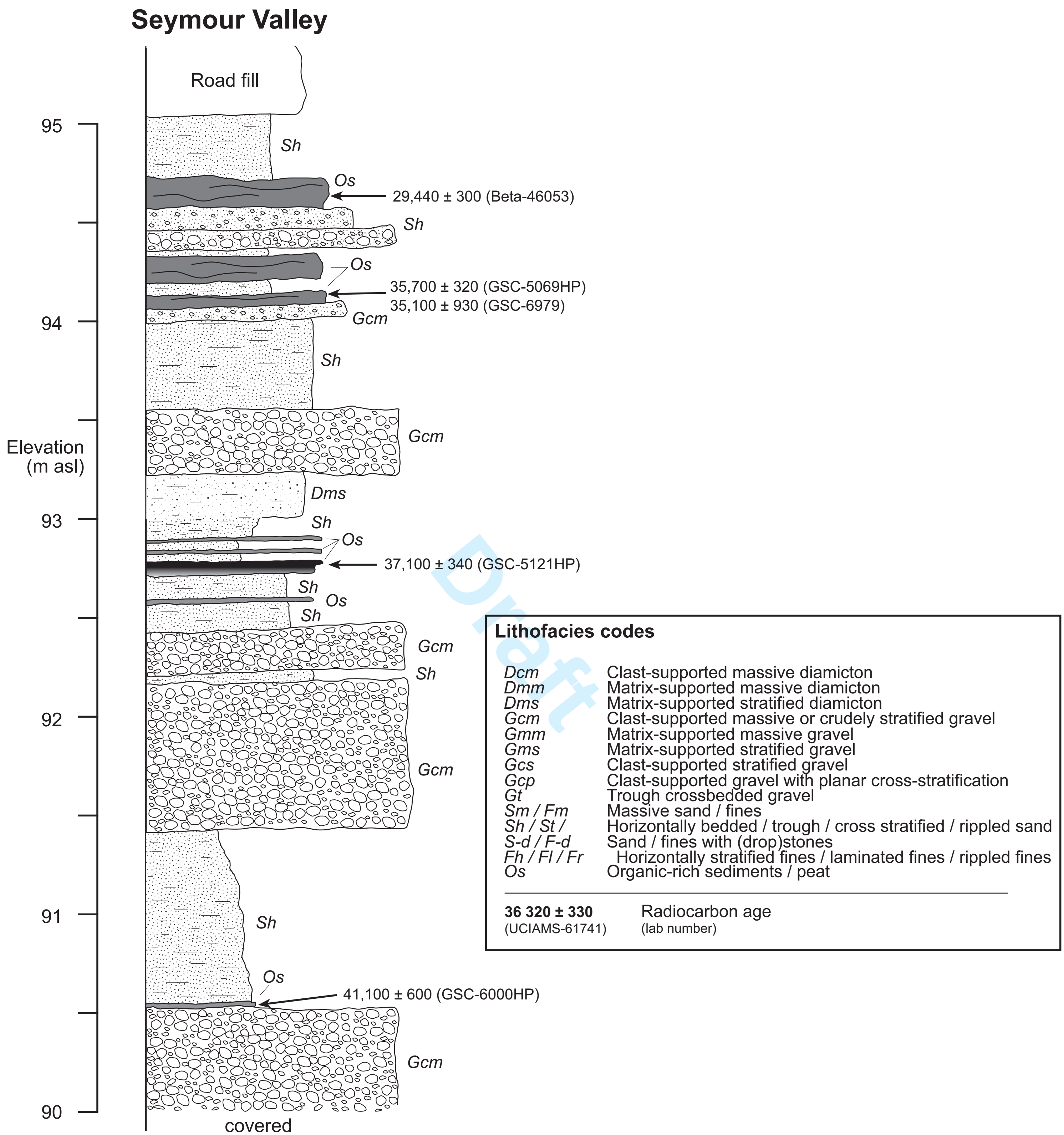
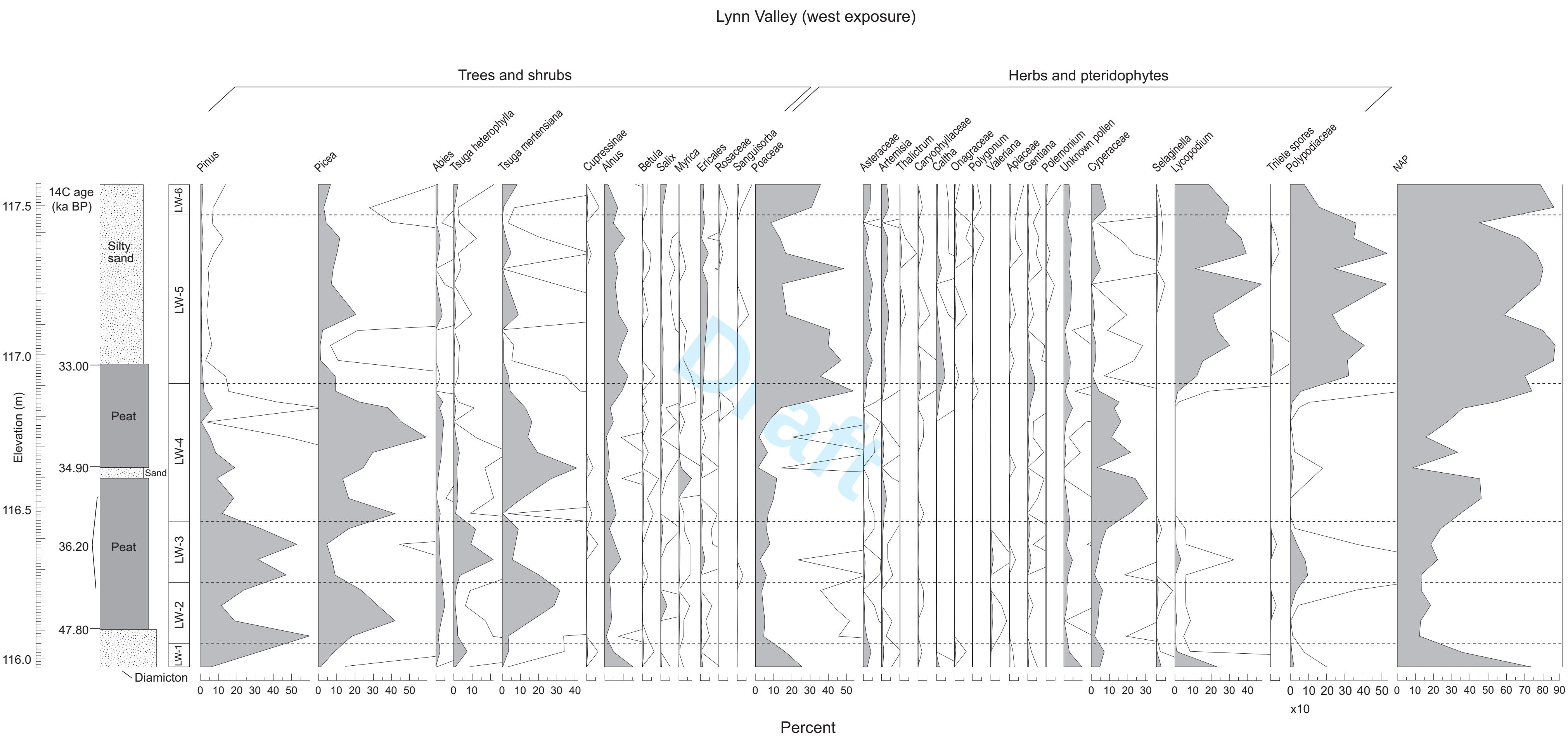
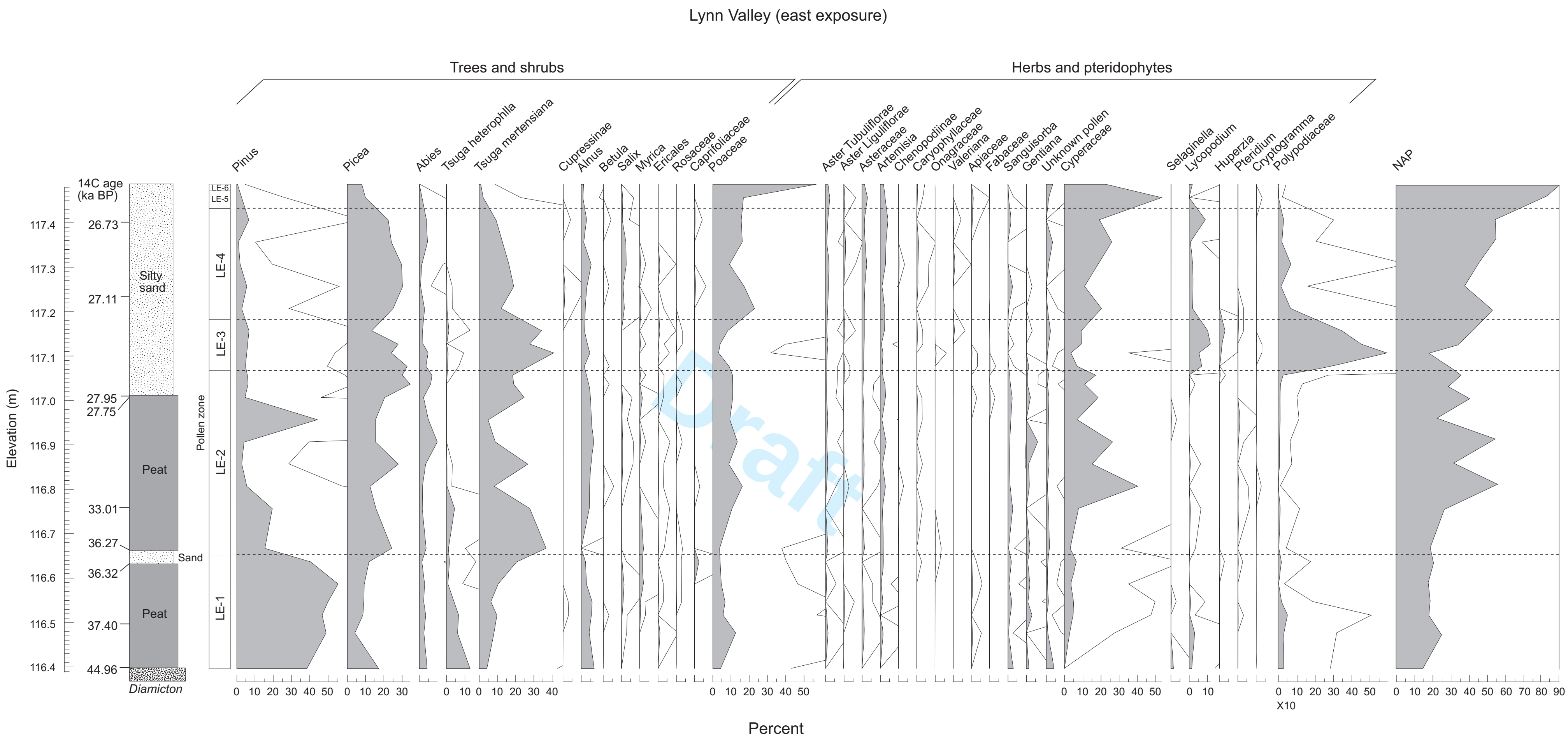


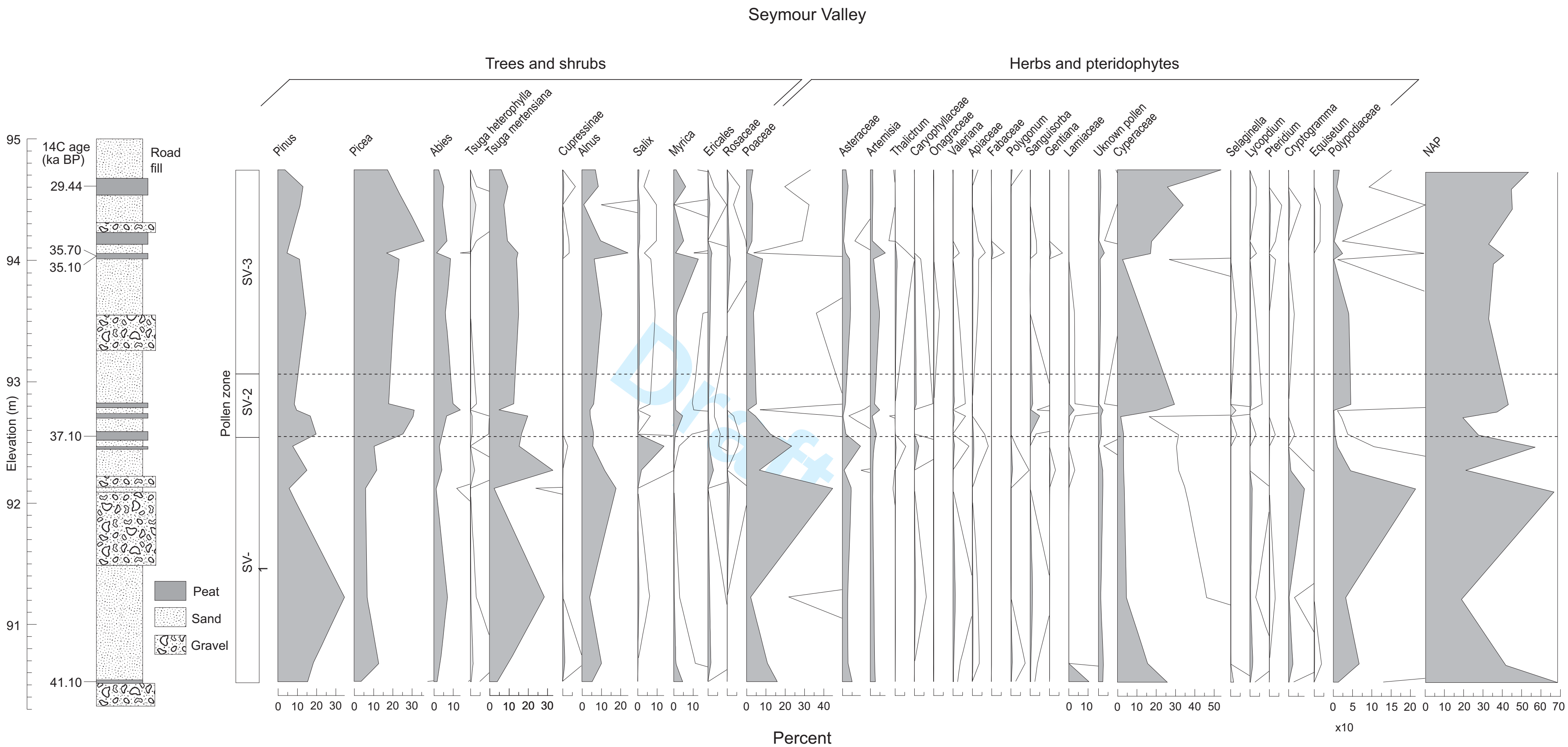

Figure 3. Lynn Valley east exposure showing the compressed peat bed interrupted by a sand bed about 5 cm thick, as it appeared in 2006. The peat bed is separated from the underlying diamicton by a unit of organic-rich colluvium likely derived from the diamicton. The peat is overlain by silty sand. The units on the scale bare are 1 cm.
469x618mm (300 x 300 DPI)


Seymour Valley
Road fill
95
Sh
Os
29,440 ± 300 (Beta-46053)
Sh
Os
35,700 ± 320 (GSC-5069HP)
35,100 ± 930 (GSC-6979)
Gcm
94
Sh
Gcm
Elevation
(m asl)
Dms
93
Sh
Os
37,100 ± 340 (GSC-5121HP)
Sh
Os
Sh
Gcm
Sh
92
Gcm
91
Sh
Os
41,100 ± 600 (GSC-6000HP)
Gcm
90
covered
Lithofacies codes
Dcm Clast-supported massive diamicton
Dmm Matrix-supported massive diamicton
Dms Matrix-supported stratified diamicton
Gcm Clast-supported massive or crudely stratified gravel
Gmm Matrix-supported massive gravel
Gms Matrix-supported stratified gravel
Gcs Clast-supported stratified gravel
Gcp Clast-supported gravel with planar cross-stratification
Gt Trough crossbedded gravel
Sm / Fm Massive sand / fines
Sh / St / Horizontally bedded / trough / cross stratified / rippled sand
S-d / F-d Sand / fines with (drop)stones
Fh / Fl / Fr Horizontally stratified fines / laminated fines / rippled fines
Os Organic-rich sediments / peat
36 320 ± 330 Radiocarbon age
(UCIAMS-61741) (lab number)

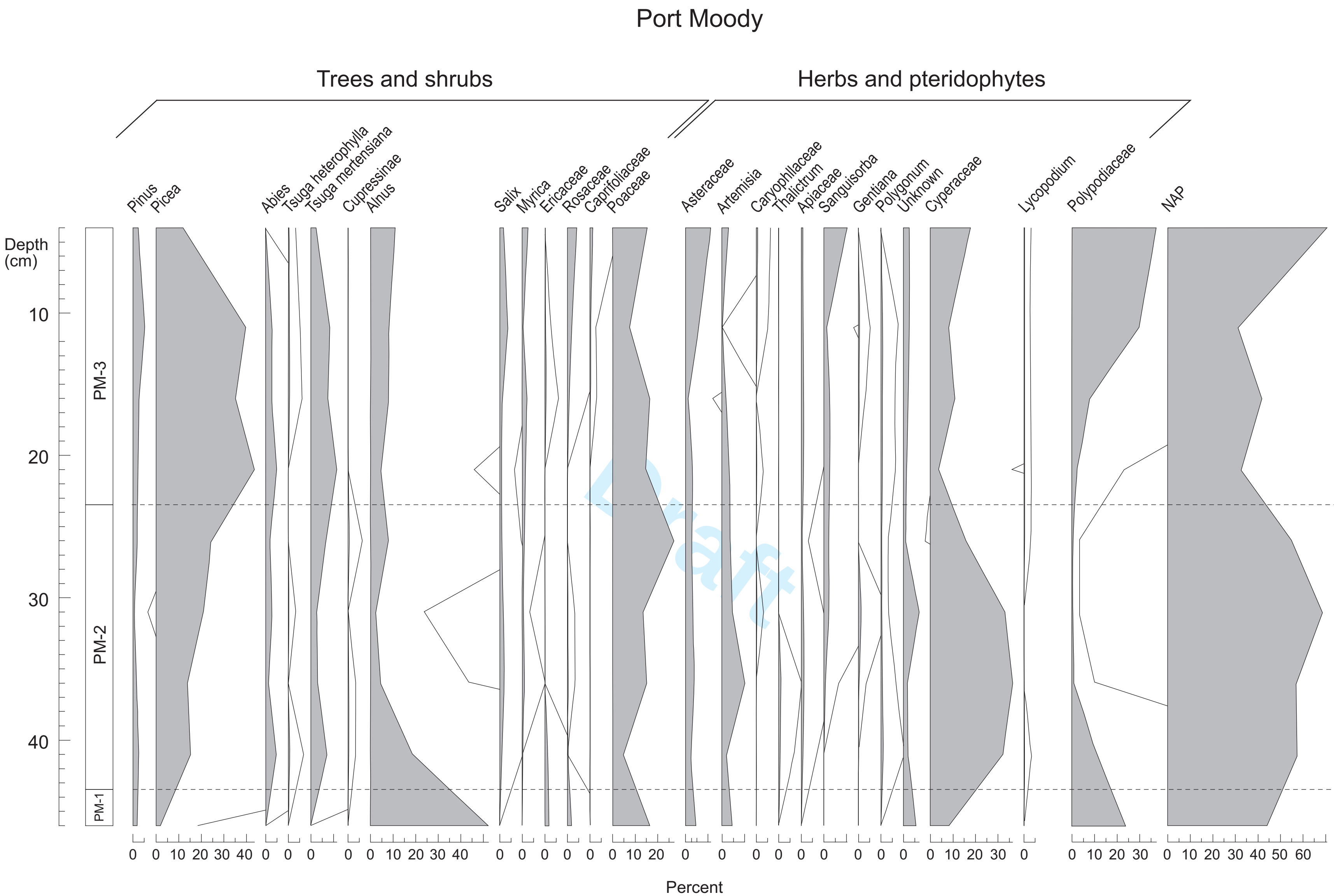


Lynn Valley (west exposure)
Trees and shrubs
Herbs and pteridophytes
Pinus
Picea
Abies
Tsuga heterophylla
Tsuga mertensiana
Cupressinae
Alnus
Betula
Salix
Myrica
Ericales
Rosaceae
Sanguisorba
Poaceae
Asteraceae
Artemisia
Thalictrum
Caryophyllaceae
Caltha
Onagraceae
Polygonum
Valeriana
Apiaceae
Gentiana
Polemonium
Unknown pollen
Cyperaceae
Selaginella
Lycopodium
Trilete spores
Polypodiaceae
NAP
14C age (ka BP)
Elevation (m)
117.5
117.0
116.5
116.0
33.00
34.90
36.20
47.80
Silty sand
Peat
Sand
Peat
Diamicton
LW-6
LW-5
LW-4
LW-3
LW-2
LW-1
x10
Percent


Lynn Valley (east exposure)
Trees and shrubs
Herbs and pteridophytes
Pinus
Picea
Abies
Tsuga heterophlla
Tsuga mertensiana
Cupressinae
Alnus
Betula
Salix
Myrica
Ericales
Rosaceae
Caprifoliaceae
Poaceae
Aster Tubuliflorae
Aster Liguliflorae
Asteraceae
Artemisia
Chenopodiinae
Caryophyllaceae
Onagraceae
Valeriana
Apiaceae
Fabaceae
Sanguisorba
Gentiana
Unknown pollen
Cyperaceae
Selaginella
Lycopodium
Huperzia
Pteridium
Cryptogramma
Polypodiaceae
NAP
14C age
(ka BP)
Elevation (m)
Silty
sand
Peat
Sand
Diamicton
Pollen zone
LE-1
LE-2
LE-3
LE-4
LE-5
LE-6
X10
Percent



Seymour Valley
Trees and shrubs
Herbs and pteridophytes
Elevation (m)
14C age (ka BP)
Road fill
29.44
35.70
35.10
37.10
41.10
Peat
Sand
Gravel
Pollen zone
SV-3
SV-2
SV-1
Pinus
Picea
Abies
Tsuga heterophylla
Tsuga mertensiana
Cupressinae
Alnus
Salix
Myrica
Ericales
Rosaceae
Poaceae
Asteraceae
Artemisia
Thalictrum
Caryophyllaceae
Onagraceae
Valeriana
Apiaceae
Fabaceae
Polygonum
Sanguisorba
Gentiana
Lamiaceae
Uknown pollen
Cyperaceae
Selaginella
Lycopodium
Pteridium
Cryptogramma
Equisetum
Polypodiaceae
NAP
x10
Percent


Port Moody
Trees and shrubs
Herbs and pteridophytes
Pinus
Picea
Abies
Tsuga heterophylla
Tsuga mertensiana
Cupressinae
Alnus
Salix
Myrica
Ericaceae
Rosaceae
Caprifoliaceae
Poaceae
Asteraceae
Artemisia
Caryophllaceae
Thalictrum
Apiaceae
Sanguisorba
Gentiana
Polygonum
Unknown
Cyperaceae
Lycopodium
Polypodiaceae
NAP
Depth (cm)
10
20
30
40
PM-3
PM-2
PM-1
Percent

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