The Lillooet terraces of Fraser River: a palaeoenvironmental enquiry[1]

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The history of Holocene fluvial terraces was investigated by stratigraphic, morphological, and palaeohydrologic methods. The terraces were formed in glaciofluvial and glaciolacustrine sediments by the Fraser River: degradation alternated with episodes of stability and aggradation. The uppermost terraces, four nonpaired surfaces, were occupied by a braided river. Vertical stability was controlled either by the water and sediment discharge of the river or by backwater effects from downstream landslides. The lower terraces, two paired surfaces, and associated fluvial sediments provide evidence for at least two cycles of aggradation and downcutting. These can be attributed to the effects of landslides that occurred a short distance downstream. Climatic variations may also have influenced terrace formation, but direct evidence is lacking. Palaeohydraulic investigations based on gravel texture, terrace gradients, and geometry of palaeochannels provide results that although approximate, conform with conclusions based upon terrace morphology and stratigraphy.

L'histoire des terrasses fluviatiles holocènes a été étudiée par les méthodes stratigraphiques, morphologiques et paléohydrologiques. Les terrasses se sont développées dans les sédiments glaciofluviatiles et glaciolacustres par la fleuve Fraser : des épisodes de stabilité et d'alluvionnement ont alternés avec des périodes d'érosion. Les terrasses sommitales, incluant quatre surfaces nonpairées, ont été envahies par une rivière anastomosée. Le contrôle de la stabilité des parois verticales était assuré soit, par l'eau et le déversement des sédiments de la rivière ou par les effets du ressac des glissements de terrain en direction avale. Les terrasses de fond, comprenant deux surfaces pairées, et les sédiments fluviatiles associés prouvent l'existence d'au moins deux cycles d'alluvionnement et d'érosion. Ces cycles sont interprétés comme le résultat probable des effets des glissements de terrain qui sont apparus à faible distance en aval. La formation des terrasses a possiblement été influencée par les changements climatiques, sans avoir laissé de preuves. L'étude paléohydraulique fondée sur la texture du gravier, les gradients des terrasses et la géométrie des paléochenaux apporte des résultats qui, bien qu'approximatifs, sont en accord avec les conclusions tirées à partir de la morphologie et de la stratigraphie.

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Introduction

Holocene river terraces occur in the Fraser River valley near the town of Lillooet, 340 km upstream from the mouth of the river (Fig. 1). Thick fluvial sediments underlie the terraces, indicating that they are not of simple degradational origin. The objective of this paper is to describe the development of the terraces, including some aspects of the hydrology of the Fraser River in Holocene time.

We first describe the morphology and stratigraphy of the Lillooet terraces, with evidence drawn chiefly from the left (east) bank of the Fraser River, where most exposures occur. Several stages may be deduced in the development of the terraces. Details of palaeochannel geometry available at several sites allow palaeohydraulic computations to be made. These serve to confirm some aspects of the interpretation of development. They also allow some characteristics of earlier Holocene flood flows to be compared with flows in the contemporary river. Substantial uncertainties remain in such a computational exercise because of the limited exposure of sediments, limited measurement possibilities, and the limits of measurement precision. These issues are examined in some detail.

Physiography and late Quaternary history

In the Lillooet area, the Fraser River is deeply incised along the fault zone that separates the Coast Mountains from the Clear Range of the Interior Plateau (Duffell and McTaggart 1952; Monger and McMillan 1984). The valley is deep and narrow, with benches of drift and nonglacial sediments standing up to 200 or 300 m above river level. Postglacial downcutting of the Fraser and many other rivers in interior British Columbia is, in general, attributable to early Holocene decline of sediment supply and subsequent reduced sediment load (see Church and Ryder 1972). At Lillooet, where the Seton River joins the Fraser River, the valley widens to about 6 km, and bedrock is more than 62 m below river level (Ryder, unpublished manuscript). This part of the Fraser Valley will be referred to as "Lillooet basin" (Fig. 2). Local widening and deepening are attributed to erosion by a glacier that flowed into the Fraser Valley from the Seton Valley, most recently during early Fraser (= late Wisconsinan) Glaciation, but probably also during earlier glaciations.

The late Quaternary history of Lillooet basin and adjacent parts of the Fraser Valley has previously been described by Ryder (1976, 1981, unpublished manuscript). Glaciolacustrine silt with minor clay and sand—"Lillooet silt" (Ryder, in press)—occupies the southern part of the basin (Fig. 3). It was deposited in a lake that was impounded north of Lytton by ice in the Fraser Canyon prior to the Fraser Glaciation maximum. Lillooet silt was partly eroded during Fraser Glaciation and then buried by glaciofluvial sediments during deglaciation 10 000 to 11 000 ^{14}C years BP (dates in Clague 1981). The Riverlands kame terrace (Figs. 2, 3) was probably deposited between two ice masses in the Fraser Valley by meltwater issuing from the Seton Valley during a pause in glacier reces-

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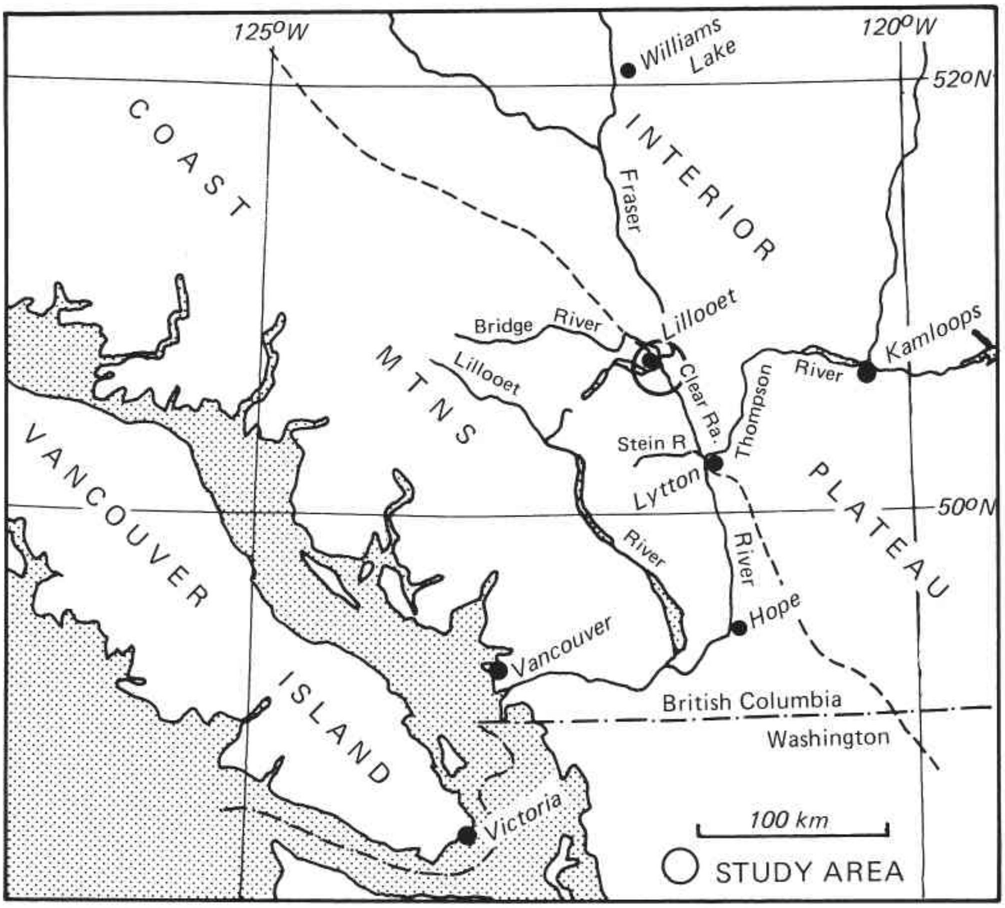

FIG. 1. Location map.

sion. Kame terraces that extend upstream from Lillooet basin to beyond the mouth of the Bridge River are more typical ice-marginal features. Ice-contact gravels outcrop in river banks, terrace scarps, and roadcuts throughout the northern part of Lillooet basin, whereas Lillooet silt is exposed to the south. A few small exposures show that the glaciofluvial gravels, particularly the Riverlands kame terrace, overlie Lillooet silt, but the overall relationship is not clearly demonstrated and is probably more complex than indicated in Fig. 3. Regularly stratified outwash gravels outcrop at a few sites, but there was no prolonged aggradation of outwash.

During Holocene time, the Fraser River has become entrenched in alternating drift and bedrock along the valley. Although Lillooet basin near the mouth of the Seton River is underlain by thick drift, bedrock is exposed in the river channel a short distance upstream and downstream. At the south end of Lillooet basin, near Jones bench (Fig. 2), the Fraser occupies a 160 m deep gorge in bedrock and slide rock.

Terrace morphology

Six morphologically distinct terraces are present between 12 and 100 m above river level in Lillooet basin (Figs. 2, 3). The upper four, T1 and T2 on the east bank and L1 and L2 on the west bank, constitute a nonpaired sequence. The two lower terrace levels, T3–L3 and T4–L4, constitute a paired sequence. The highest terraces are most extensive, and lower terraces are relatively narrow; T4 and L4 are restricted to the southern part of the study area. The modern channel of the Fraser River is confined between bluffs of drift and bedrock, with virtually no floodplain.

Although strikingly regular when viewed from a distance (Fig. 4*a*), the terraces are irregular in detail. Aeolian veneers of variable thickness and dunes that have been smoothed in cultivated areas give rise to gently undulating topography with local relief of 2 or 3 m. The continuity of some of the main terrace surfaces is interrupted by scarps of up to about 7 m relief. T3 in particular is broken into several subunits; one of these scarps coincides with the only bedrock outcrop that was found on the terraces. Scarps and abandoned channels masked by aeolian sediments may also constitute terrace surface irregularities. Parts of some surfaces, particularly T3, have been extensively modified and probably slightly lowered by placer mining.

Abandoned channels are not well preserved on the terraces. One channel is reflected by topography on T2, and traces of several small channels, probably part of a braided system, are detectable on T1.

Measurements of terrace elevations and longitudinal gradients were obtained from large-scale topographic maps and photogrammetry. Field surveys were not successful because of the problem of spotting representative positions on the irregular terrace surfaces. Topographic maps at a scale of 1 : 6 000 with 40 ft (12.2 m) contour interval (British Columbia Department of Lands and Forests 1956) were used to interpolate elevations, which are considered to fall within ± 10 ft (± 3 m) precision (Table 1; Figs. 3, 5). Terrace gradients were estimated from elevations determined photogrammetrically in a

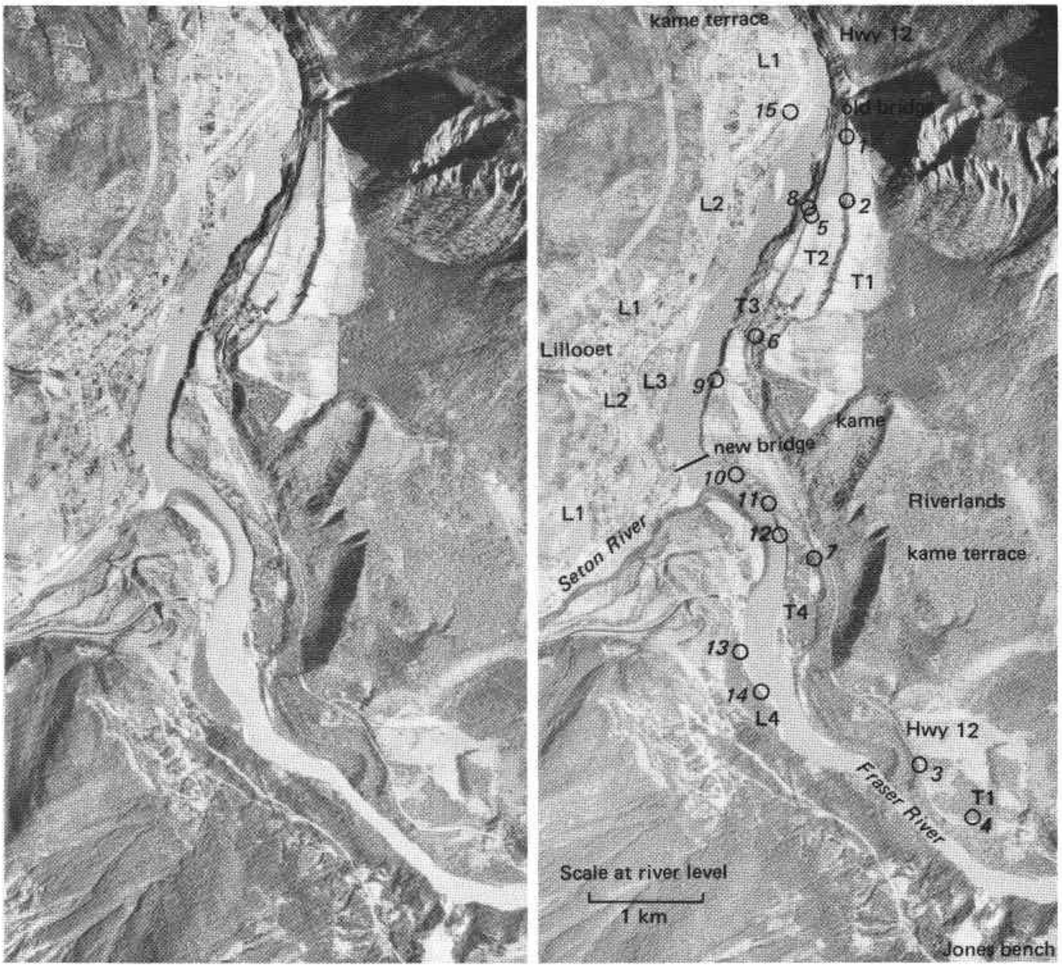

FIG. 2. Stereopair of Lillooet basin: morphological features and locations of sections (circles).

Wild A6 plotter. Individual surface points were estimated to within ± 0.5 m; however, terrace surface elevations are considered to be estimated only to within ± 1.0 m for gradient determinations because of the variable aeolian veneer.

Longitudinal gradients estimated for the terraces range from 4.2×10^{-3} to 2.6×10^{-3} and are steeper than the present gradient of the Fraser River (1.7×10^{-3}). Transverse gradients on most of the terraces are relatively small, generally less than 1° (20×10^{-3}), although in a few places, most notably on L3 within Lillooet townsite, much steeper cross slopes exist locally where colluvium derived from an adjacent valley side or uphill scarp has been deposited on the terrace.

As far as is known, there has been no modification of terrace gradients by tectonic or isostatic processes. The most recent displacements along the Fraser River fault zone are pre-Quaternary (see Mathews and Rouse 1984). Fulton (1969) demonstrated that isostatic tilting of the southern Interior Plateau, which adjoins our study area, ceased before 9000 years BP.

Well-developed terraces like those of Lillooet basin are not present in adjoining parts of the Fraser Valley. Downstream, between Lillooet basin and the mouth of the Stein River and for about 40 km south of Lytton, the upper surface of the Quaternary valley fill is an irregular and discontinuous bench, commonly with a strong transverse slope on large alluvial fans of early Holocene age. Only in one or two places does this bench include ''level'' areas that may be river terraces equivalent to the uppermost (T1) terrace at Lillooet. Narrow fluvial terraces are present at many different elevations between the main bench and the river; however, their extent is so small and their surface morphology has been so modified by aeolian material and colluvium that with the exception of a few terraces close to present river level, they cannot be interrelated or correlated with the Lillooet terraces. Upstream, the Fraser River occupies a deep gorge that is flanked by high kame terraces and fan-covered benches of Quaternary sediments and bedrock; river terraces are absent (Ryder 1976).

Well-developed river terraces that are broadly similar to those of Lillooet basin are present in the wide, 6 km stretch of Fraser Valley between the Stein River and Lytton (Ryder 1981), in the lower Bridge River valley (Howes 1975), and in the lowermost Seton River valley (Fig. 2).

Terrace sediments

Holocene fluvial sediments and associated materials were examined in all exposed sections in the east bank terraces (T1 – T4) and briefly in the west bank terraces (L1 – L4) (Figs. 2, 4*b*, 5). T3 and T4 sediments are well exposed in cutbanks that extend continuously for about 1.5 km along the east bank of the Fraser River near the mouth of the Seton River (sections 10, 11, 12). Exposures in T2 and T3 are chiefly roadcuts 100 – 200 m long. The fluvial sediments comprise two distinct facies: overbank sediments and channel gravels.

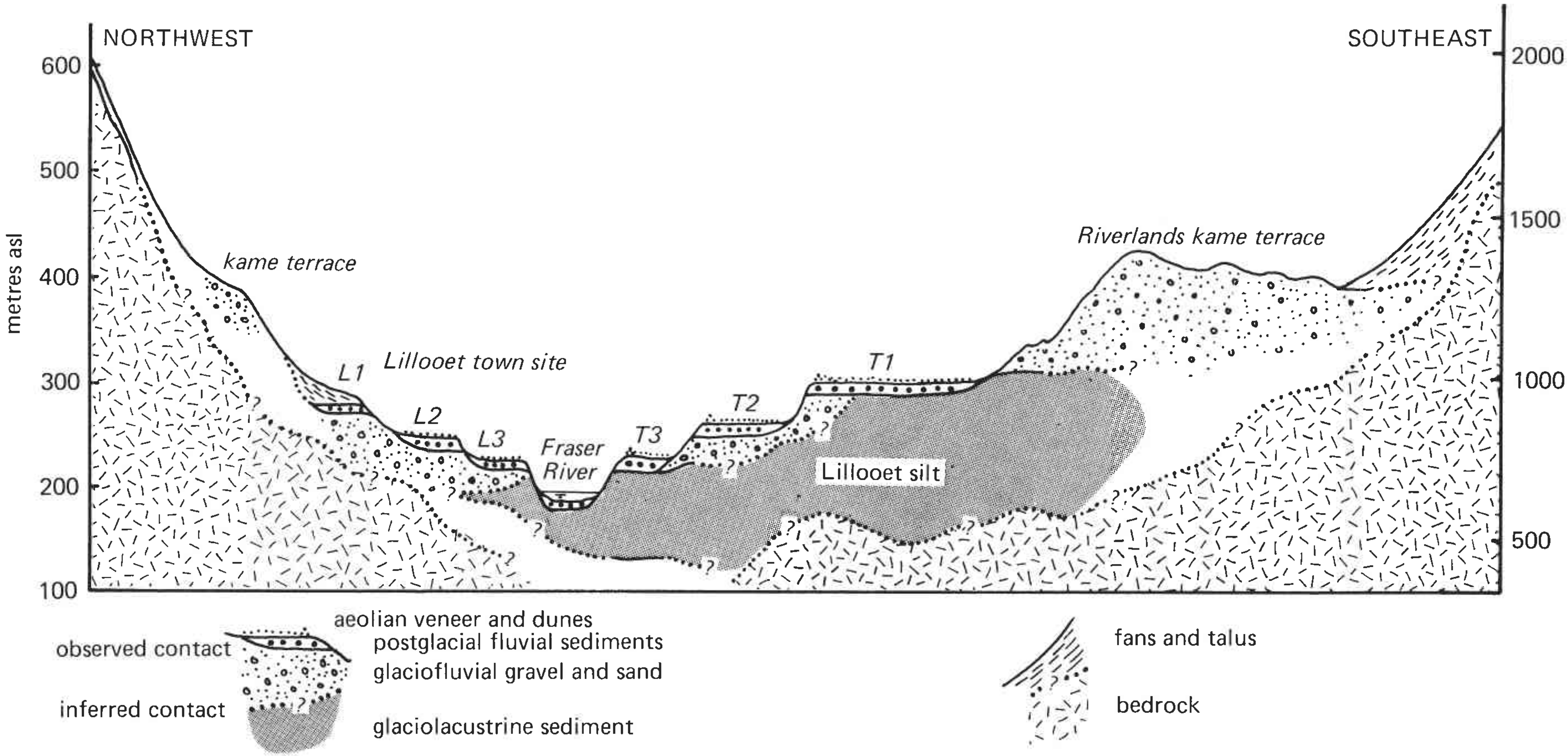

FIG. 3. Schematic cross-section from northwest to southeast across Lillooet basin, showing the relationship between terraces, fluvial sediments, and drift units. (Thicknesses of fluvial and aeolian sediments are not to scale; the lowest terraces are not shown.)

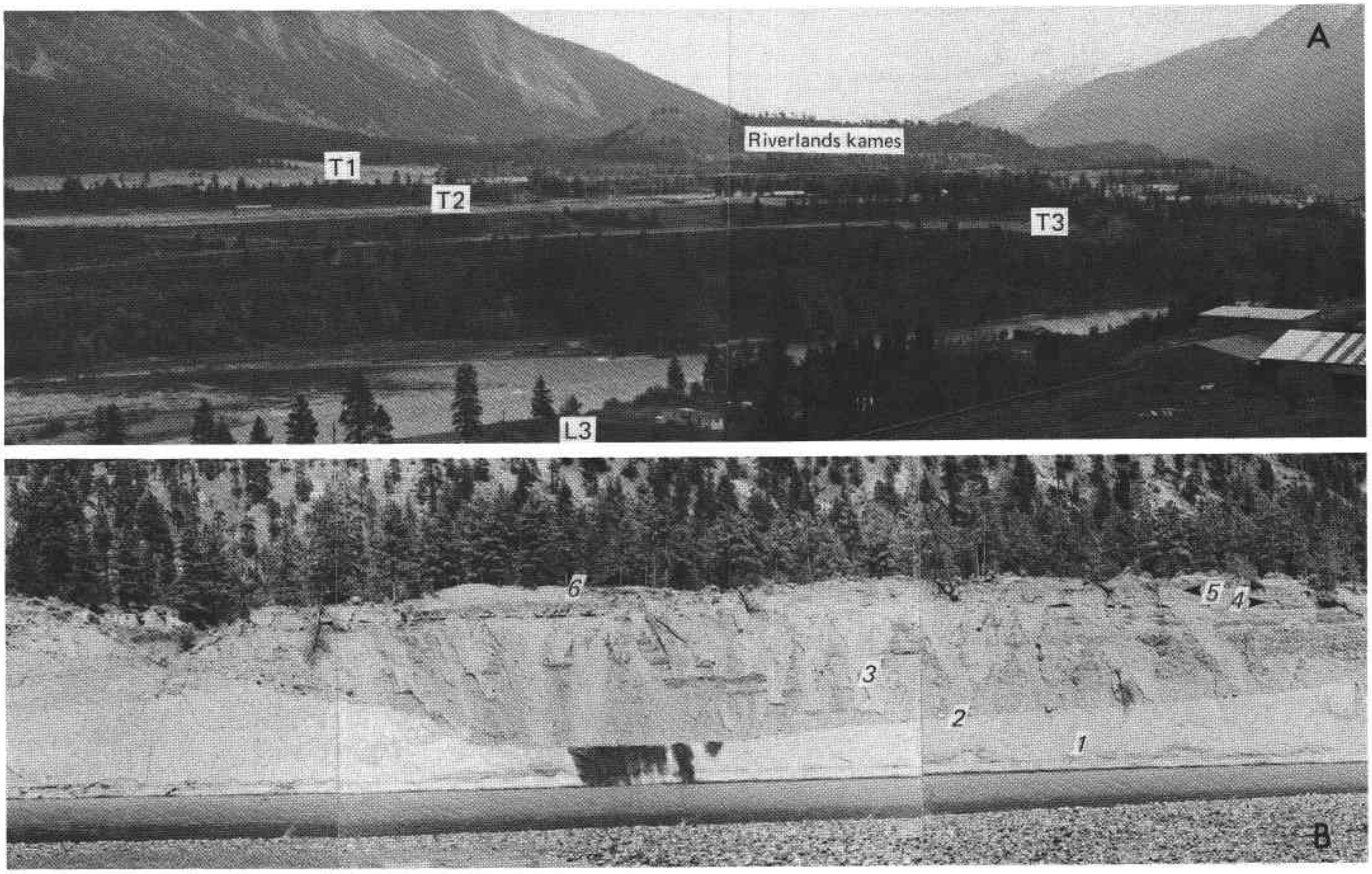

FIG. 4. (*A*) View southeastward across Lillooet basin. (*B*) Infilled channel below T3 at section 11 (360–550 m in Fig. 6): (1) Lillooet silt; (2) lag gravel (here a row of boulders); (3) stratified gravel; (4) overbank sediments; (5) aeolian veneer; (6) dunes.

TABLE 1. Terrace elevations and gradients

	Elevation[a]		Gradient ($\times 10^{-3}$)	
Terrace	m above sealevel	m above Fraser River	from photogrammetry	from map[b]
T1	295	103	2.6 ± 0.2	2.8 ± 0.4
L1	275	83		
T2	260	68	3.6 ± 0.6	3.2 ± 0.6
L2	245	53	4.7 ± 1.0	3.3
T3	230	38	4.2 ± 0.8	4.3 max
L3	230	38	3.3 ± 1.5	
T4	200	8		
L4	200	8		
Fraser River	192	—	2.3 ± 0.5	1.52[c] (low flow) 1.72[c] (high flow)

[a]On an east–west transect of Lillooet basin 0.8 km north of the mouth of the Seton River.

[b]From large-scale topographic maps with 40 ft (12.2 m) contour intervals. The map contours were assumed to be correct, and the error measures represent interpolation error. Exact numbers are measured from contour to contour. Actual precision would be lower.

[c]Surveyed over 2.3 km downstream from the new bridge. The photogrammetric estimate includes a steeper reach immediately upstream.

Overbank sediments

These consist of thinly interbedded, commonly laminated, well-sorted silt and sand, most typically coarse silt and fine sand. Thin beds of granule gravel or pebbly sand are included within the finer sediments at some sites. Bedding is typically well developed and clearly visible from a distance, which enables fine fluvial sediments to be distinguished readily from aeolian deposits of similar texture. Normally, bedding is horizontal, but it may be gently inclined where overbank sediments overlie minor terrace scarps or fill shallow channels. The thickness of overbank sediments ranges from a few centimetres to about 3 m.

Channel gravels

These may be divided into two subfacies: boulder (including lag) gravels and stratified gravels.

Boulder gravels

Such gravels constitute the coarsest stratigraphic unit. They typically consist of a massive bed of rounded boulders and large cobbles supported by a loosely packed matrix of smaller, less rounded clasts and sand. The largest boulders are generally about 500 mm in diameter. Bed thickness varies from a maximum of 2 m to the diameter of a single clast. Relatively thick beds may include lenses of framework-supported boulders and cobbles. The larger clasts, which include many platy boulders, show strong imbrication. Lag gravels form boulder and cobble pavements, which appear in sections as continuous rows of large clasts and as individual large clasts spaced at intervals along a particular stratigraphic horizon.

Boulder gravels are normally, although not invariably, present at the erosional contact between fluvial sediments and underlying drift (Fig. 5). Bouldery gravels are most common on top of glaciofluvial gravels, whereas individual large clasts predominate on top of Lillooet silt. Large, angular blocks

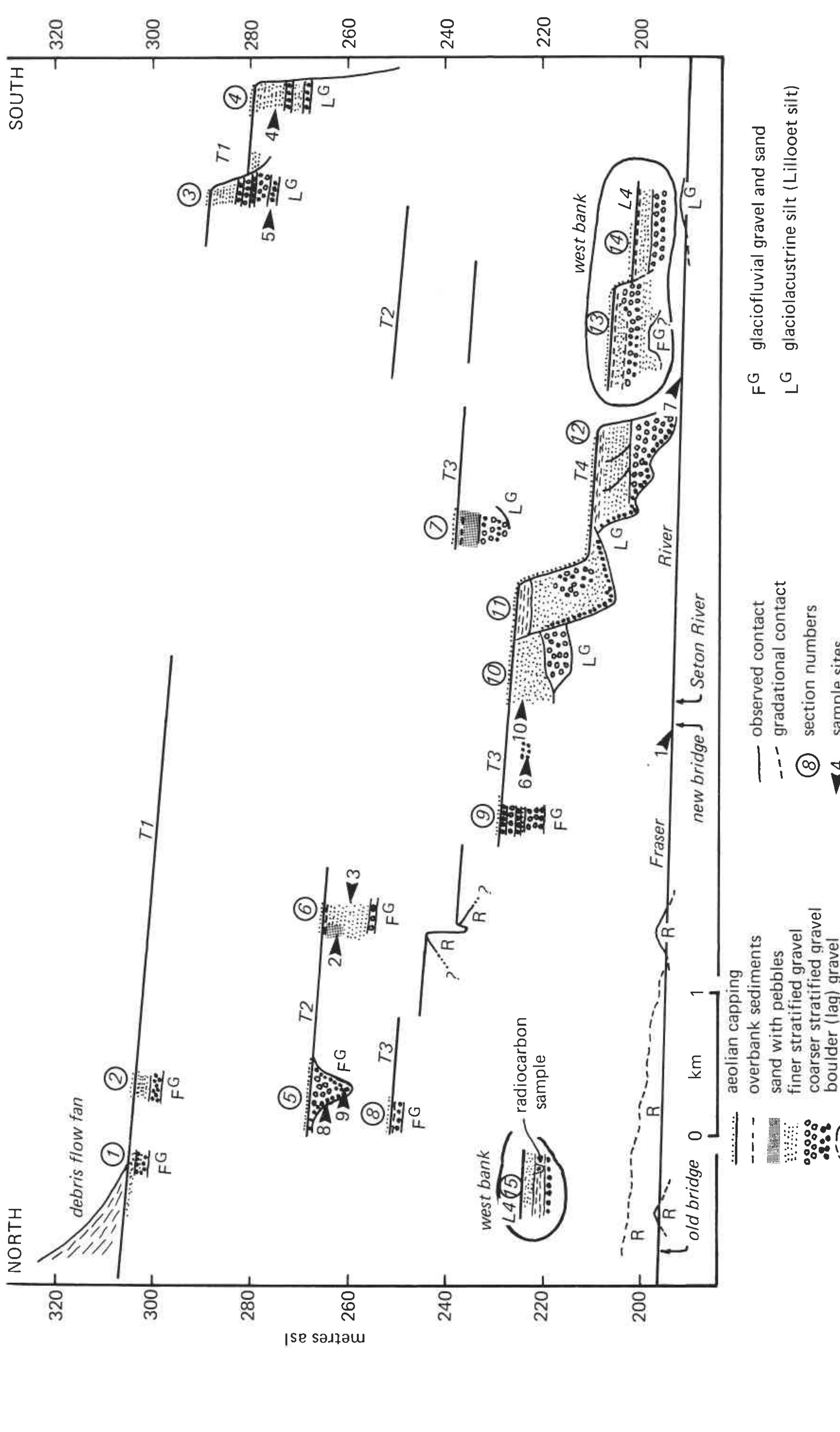

FIG. 5. Schematic drawing of terraces and measured sections on the east bank. Stratigraphy of west bank sections is from field sketches. Absolute elevations, gradients, and horizontal scale are approximate. Section locations are shown in Fig. 2.

(1–2 m in diameter) derived from local bedrock are present along the contact in some sections. Beds of bouldery gravel and lag pavements also occur within stratified gravels.

Boulder gravels are concentrations of large clasts that accumulated along the deepest parts of the river channel (thalweg) and on bar heads during local or general degradation. This material (except lag gravels) was transported during high flows. Whether deposition resulted in framework- or matrix-supported boulder gravel probably depended upon the relative proportions of boulders and smaller clasts that were mobile and upon details of the flow. Smaller clasts in framework-supported beds may have been trapped between boulders during deposition or washed in at a later time.

Stratified gravels

These are moderately well sorted gravels consisting of subangular to rounded cobbles, pebbles, granules, and sand. They are interpreted as bar gravel accumulations. Generally, stratification is horizontal and is typical of the sheet-like accretion of bed material onto gravel bar surfaces that occurs in large channels. Cross-bedding is rare. The thickness of stratified gravels is typically between 2 and 10 m, with a maximum of 15 m (Figs. 4*b*, 5).

In many sections, there are distinct units of uniform, relatively coarse and relatively fine stratified gravels, each several metres in thickness. Most commonly, an upper unit of sandy pebble gravel is separated by a sharp horizontal contact from underlying poorly sorted cobble gravel (Fig. 5). The upper pebble gravel may represent local aggradation during channel abandonment and filling or general aggradation in backwater.

Aeolian sediments

Fluvial materials are overlain by veneers and dunes of aeolian sediment derived from the active channel zone of the Fraser River and from unvegetated riverside bluffs of drift and fluvial materials. Aeolian veneers normally consist of massive, silty, fine sand, but there are vertical gradations of texture to coarser and finer materials that reflect increased or decreased proximity, vertically and (or) horizontally, of the Fraser River. Dunes consist of moderately to well-sorted fine to medium sand with rarely exposed cross-bedding and ripple laminae.

Stratigraphic relationships and interpretation

The stratigraphy and structural features of the fluvial sediments in the Lillooet terraces are schematically indicated in Fig. 5. Fluvial sediments of from less than a metre to more than 19 m in measured thickness unconformably overlie glaciolacustrine silt and glaciofluvial gravels and sands along contacts that are surprisingly irregular in places. The most typical stratigraphic sequence is as follows:

- — aeolian veneer and (or) dunes,
- — overbank silt and sand,
- — relatively fine stratified gravel and (or) sand,
- — relatively coarse stratified gravel,
- — boulder lag gravel, and
- — drift.

Not all units are represented at every site. The basal lag or boulder gravel is almost invariably present and may be the only fluvial facies. At sections 8 and 15 boulder gravel is directly overlain by overbank sediments, but in all other sections stratified gravels constitute most of the fluvial material. In some sections, beds or lenses of boulder or lag gravel are interlayered with stratified gravels. Overbank sediments are thickest and most extensive on the lowest terraces (T4, L4) and were rarely found on T1.

Well-defined channel cross sections cut into drift are exposed in sections 5 and 11 (see Figs. 4*b* and 6—350–630 m—for the section 11 channel). At both of these sites, faint channel traces or low scarps on the terrace surface indicate that the section is oblique. The large channel at section 11 is floored by lag gravel, infilled by well-stratified gravels, and capped by relatively thick overbank sediments. The smaller channel at section 5 is about 96 m wide (true width) and is filled with poorly stratified boulder and cobble gravel. Another similarly small channel is indicated by surface topography on L1. Long riverbank exposures of stratified gravels with uniform thickness and texture, such as at sections 10, 12, 13, and 14, are probably longitudinal channel sections.

Evidence for minor aggradation (1–2 m), perhaps local, is apparent at several sections in T1, T3, and T4, where boulder gravels overlie stratified gravels (Figs. 5, 6). For example, in section 9 the sequence

- — 1.0 m boulder gravel (top),
- — 2.0 m coarse stratified gravel,
- — 0.3 m lag of boulders and large cobbles,
- — 0.7 m cross-bedded, silt-coated pebble gravel, and
- — 6 m stratified coarse gravel

indicates that on two occasions the main river channel became superimposed on former gravel bars.

Evidence for aggradation that was probably more widespread is present within the lower terraces (T3–L3 and T4–L4). At section 12, two poorly defined, shallow channels are exposed at the base of the fluvial sediments (640–810 m in Fig. 6), whilst two partly preserved (left banks only), younger channels are indicated at a higher stratigraphic level (at 860 and 900 m in Fig. 6) by inclined bedding and abrupt changes in gravel texture. Evidence of general aggradation is also present in two of the unmeasured west bank sites that are shown in Fig. 5. At section 13 in L4, there is an upward gradation from fine to coarse stratified gravels; overbank sediments that overlie the coarse gravels are buried by second units of fine gravel and overbank sediment, suggesting two phases of aggradation.

At site 15 on L4, overbank sediments that contain cultural material (British Columbia archaeological site EeR1-12) and palaeosols in their upper part are overlain by 2 m of fine stratified gravels, suggesting that human occupation of a relatively stable floodplain or terrace was disturbed by renewed flooding and gravel deposition.

Two additional lines of evidence suggest that downcutting and terrace formation alternated with phases of aggradation. First, available exposures indicate that fluvial sediments become thicker downstream (for example, gravels under T1 and T3 illustrated in Fig. 5). This pattern is interpreted to be the result of gravel aggradation associated with a decrease in the longitudinal gradient of the Fraser River (backwater effect) in and adjacent to Lillooet basin. Second, the paired lower terraces (T3–L3 and T4–L4) and the relatively great thickness of the gravels that lie beneath them indicate that they probably formed during (at least) two cycles of aggradation and downcutting.

These interpretations of general aggradation in Lillooet basin suggest that river behaviour may have been influenced by blockage of the Fraser River downstream. At Jones bench (Fig. 2) the Fraser River has cut a canyon through debris of a late-glacial landslide (Wick Creek slide, Ryder 1976). At least

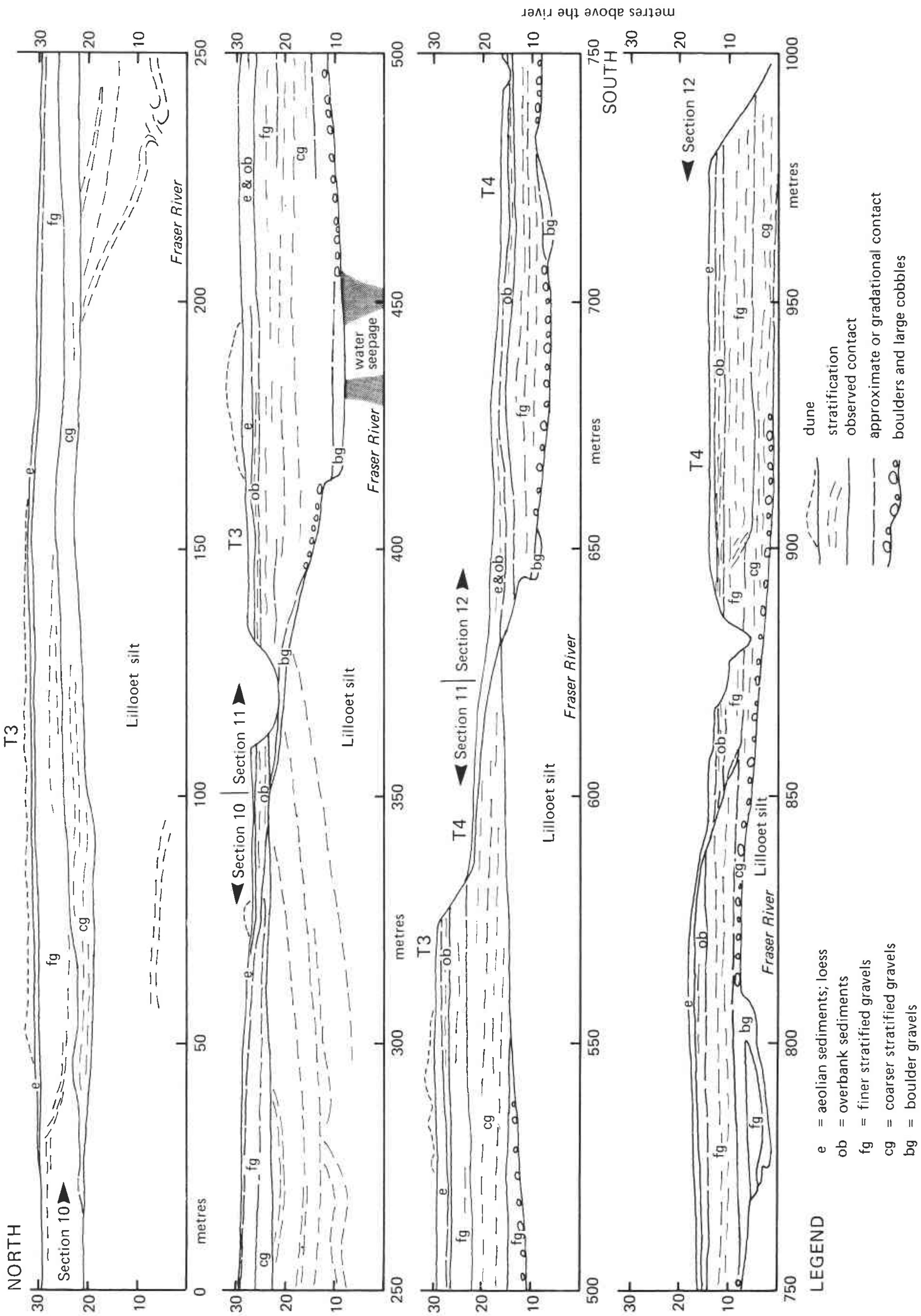

FIG. 6. Detailed scale drawing, from photographs, of the major east bank section in T3 and T4 (includes sections 10, 11, and 12 of Fig. 5).

one secondary slide has occurred here in postglacial time, and there may have been others. At Texas Creek, 17 km downstream from Lillooet, a major rockslide occurred after substantial postglacial dissection and terracing of Pleistocene valley fill had been achieved by the Fraser River and Texas Creek Ryder 1976). The presence of slide debris on a low terrace in the canyon that has been cut by the Fraser River through the debris of the initial slide, as well as interlayered river gravels and slide debris exposed in the walls of this canyon, suggests that at least two further slides occurred here. In fact, a steep, rocky slope on the eastern side of the Fraser Valley is still unstable and prohibits widening of Highway 12. A temporary dam could well be formed by a rockslide once again.

The general elevation of the crest of the major landslide dam at Texas Creek, about 305 m, is sufficiently high that the initial slide could have produced backwater in the Fraser River at the level of any of the Lillooet terraces (see Fig. 5). The known secondary slides at Texas Creek and Jones bench, with general debris elevations of about 220 m, could well have affected terraces T3–L3 and T4–L4.

Age of the terraces

Direct evidence about the age of the Lillooet terraces is available from only two sites. At section 4 in T1, Mazama tephra (6700–7000 ^{14}C years BP; Sarna-Wojcicki *et al* 1983) rests upon thin (20 cm) overbank sediments and is overlain by 1.25 m of aeolian silty sand. This provides a minimum date for the abandonment of this terrace surface by the Fraser River. If it is assumed that aeolian deposition on the terrace surface began shortly after fluvial activity ceased, then terrace formation about 7000 ^{14}C years BP is implied. Observations along the modern floodplain suggest that exposed fluvial surfaces are indeed rapidly covered by windblown sand, even during the course of a single low-water season. Prompt burial of Mazama ash by aeolian material is implied by its preservation. Consequently, a 7000 ^{14}C year age for this terrace is tentatively accepted. Stratified gravels with interlayered boulder beds at sites 3 and 4 (Fig. 5) suggest that the two terrace levels at the south end of T1 resulted from two phases of local aggradation. It is the most recent of these phases that apparently ended about 7000 ^{14}C years BP.

Little is known about the absolute age of fluvial terraces elsewhere in the Fraser–Thompson system, but comparison with the few dated localities suggests that 7000 years is relatively young for a terrace 90 m above river level. Archaeological studies at Nesikep Creek, 25 km south of Lillooet, showed that a terrace that is presently 65 m above river level was close to the river more than 6000 ^{14}C years ago, when it was occupied by palaeoindians (Sanger 1970). In the Thompson Valley at Spatsum, about 40 km upstream from Lytton, the presence of Mazama ash in alluvial fans indicates that the 40 m terrace upon which they lie is older than 7000 ^{14}C years (Ryder 1976). At Drynoch, about 15 km upstream from Lytton, cultural remains that underlie Mazama ash on the 40 m terrace date from about 7500 ^{14}C years BP (Lowden *et al.* 1969).

At section 15 in L4 (Fig. 5), charcoal from storage pits that were excavated in overbank sediments by palaeoindians (A. Stryd, personal communication, 1985) provided a date of 1180 ± 110 ^{14}C years BP (GSC-4101). This is a minimum age for the buried terrace surface that is represented by the top of the overbank sediments, and more significantly, it is also a maximum date for burial of the terrace (and cultural features) by Fraser River gravels. Thus 2 m of aggradation (at site 15) followed by 28 m of fluvial downcutting (relative elevations here were measured by altimeter survey) has occurred during the past 1200 ^{14}C years.

If aggradation was triggered as a result of damming by a landslide downstream, gravel deposition could have occurred very rapidly, probably during only a few years. The downcutting that followed, 28 m in a maximum of about 1200 ^{14}C years, appears to have been rapid, at rates of at least 2.2–2.6 cm/^{14}C year (the range derives from the standard deviation of the ^{14}C date). However, mean rates of longer term downcutting at section 4 (90 m in 7000 ^{14}C years, for 1.3 cm/^{14}C year) and at Nesikep Creek (65 m in 6000–7000 ^{14}C years, for 1.1–0.9 cm/^{14}C year) are broadly comparable, especially inasmuch as the section 4 estimate represents only a minimum value, since downcutting of the Fraser River was interrupted by at least two phases of aggradation (i.e., construction of T3–L3 and T4–L4; see below).

Summary of terrace history

Evidence described above indicates that terrace formation and fluvial sediment deposition by the Fraser River in Lillooet basin were episodic and not simply the result of continuous, gradual degradation. The relatively broad uppermost terrace surface, T1, may have originated during deglaciation as an outwash plain and then remained "active" for several millenia until commencement of general fluvial downcutting at the relatively late date of about 7000 ^{14}C years BP. Postglacial degradation along the Fraser Valley in the vicinity of Lillooet basin may have been retarded in early Holocene time because of the maintenance of high local base levels along the Fraser River by a rockslide and bedrock at Jones bench, a rockslide at Texas Creek, and aggrading alluvial fans between Texas Creek and Lytton. Alternatively, high sediment loads in the Fraser River, derived from the adjacent Coast Mountains (e.g., via Bridge River) or from erosion of drift upstream, may have precluded degradation. It is also possible that downcutting commenced when the onset of cooler and moister climatic conditions, 6000–7000 years ago at the end of the early Holocene Xerothermic interval (see Mathewes (1984) and references therein), resulted in increased discharges in the Fraser River.

Multiple channel traces on T1 and the small size of preserved channels on T2 and L1 suggest that the river that occupied these upper terraces was braided. The absence of overbank sediments on T1 is consistent with continual reworking of the terrace surface by shifting channels. Whether downcutting between terraces was effected by a single thread channel or whether a braided channel was maintained during degradation cannot be determined from available evidence, although decreasing terrace width at lower elevations suggests that the degrading river was increasingly confined between bluffs of drift. In general, the nonpaired nature of the upper terraces (T1, T2, L1, L2) indicates that the preserved surfaces were most probably formed during continuous degradation.

Characteristics of the lower terraces strongly indicate that their formation was controlled by backwater effects. The early development of T3 was achieved by a laterally migrating and gradually degrading channel. This formed the stepped, upstream part of the preserved terrace and the large channel at section 11 cut into lacustrine silt (Fig. 4*b*), which is similar in size to the modern Fraser channel. Confinement by the silt may have effected the pattern transition to single thread.

The onset of backwater conditions, apparently controlled by landslide dams downstream, resulted in gravel aggradation to the main T3 level. This event is possibly marked by the abrupt upward change from coarser to finer stratified gravel. A period of stability ensued, during which substantial overbank sediments accumulated before renewed downcutting resulted in the accumulation of boulder and lag gravels at stratigraphic levels above the drift – fluvial sediment contact and in the development of the paired T3–L3 terrace. Terraces T4–L4 were formed by one or more repetitions of a similar cycle of events.

Much of the foregoing summary is necessarily deductive, in view of the sporadic preservation (or exposure) of evidence. Some further confidence would be lent to the interpretations if they could be shown to be in accordance with morphological requirements of flows in the Fraser River. Three particular questions may be pursued. Were the flows on T1 and T2 apt to occur in braided channels? Were the flows in T3–T4 time greater or smaller than those in the present river? Can the supposed backwater sediments be confirmed as such from hydraulic arguments?

Palaeohydraulics

It is possible to estimate the magnitude of the flows that deposited the fluvial materials underlying the Lillooet basin terraces if certain field measurements are available and if certain assumptions are accepted. Field data of channel gradient and bed grain size can be used in formulae for critical tractive force (to move the grains) and flow resistance to estimate flow velocity and specific discharge. If channel sectional geometry is also available, the total discharge may be estimated. This can be accomplished for sites associated with T2 and T3 where channel sections are preserved. These results, along with specific discharge estimates for the three upper terraces, provide some insight into the questions posed at the end of the last section.

Texture analyses were obtained for channel gravels in exposures in each of terraces T1, T2, and T3 and from modern river bars. Sampling criteria, procedures, and results are given in Appendix 1. Grain-size parameters for analyses were obtained from texture curves.

We begin from Shields' criterion for the shearing stress necessary to initiate movement of a particle on the bed of a stream:

$$[1]\quad \tau^*_{ci} = \tau_{ci}/s'_s D_i$$

where τ^*_{ci} is the dimensionless shear to move a grain of diameter D_i; τ_{ci} is the dimensioned shear (kg m^{-2}); and s'_s is the submerged specific weight of the particle, usually taken to be 1650 kg m^{-3}. The usually quoted values of τ^*_{ci} fall within the range of $0.045 \leq \tau^*_{ci} \leq 0.060$, but higher values may be assigned if the particle is initially constrained by its neighbours, that is, if it is imbricated (see Church (1978) for discussion). However, we are concerned here with the cessation of motion of particles that will certainly have been unconstrained prior to lodgement. Further, we are concerned with the largest particles, since these indicate the competence of the stream at the largest flows. (Little may be said about lesser flows in the present context.) Andrews (1983), following Fenton and Abbot (1977), pointed out that large grains are prominently exposed in the flow and may be expected to move at relatively low values of τ^*_{ci}. Andrews computed the relation

$$[2]\quad \tau^*_{ci} = 0.0834(D_i/D_{b50})^{-0.87}$$

wherein D_{b50} is the median grain size of the bulk material (i.e., subsurface material); $0.3D_{b50} \leq D_i \leq 4.2D_{b50}$. In the limit $D_i \geq 4.2D_{b50}$, $\tau^*_{ci} = 0.020$. The relation is based upon observations of sediment in transport in contemporary gravel bed rivers. Hence, although Andrews entitled his paper *Entrainment of Gravel . . .*, the conditions that he studied are more analogous to the ones we wish to compute, which may be quite different from those that actually occur at entrainment. In this paper, D_i is taken as D_{90} of the subsurface material to represent the large material present in the flow. From these considerations, we obtain values of shear stress, τ_{ci} tabulated in Table 2, thence flow depth $d = \tau_{ci}/\gamma S$, where γ is the specific weight of water and S is stream gradient. Estimates of the errors associated with the observations and calculations are given in Appendix 2.

To estimate velocity we require a flow resistance relation and a means to estimate the resistance coefficient. At high flows over gravel (without primary bed forms), the various logarithmic and power law flow resistance formulae converge to the particle-Manning equation, so-called when the Manning resistance number is calculated from bed grain size. An early empirical relation, from Strickler (see Henderson 1966), was

$$[3]\quad n = 0.038D_{90}^{1/6}; \qquad D_{90} \geq 8 \text{ mm}$$

Limerinos (1970) presented the rational relation

$$[4]\quad \frac{n}{R^{1/6}} = \frac{0.113}{1.16 + 2.0 \log (R/D_{84})}$$

where $R \sim d$ is the hydraulic radius. From data in Table 2 we obtain, using [3], $0.025 \leq n \leq 0.032$ for boulder gravels and $0.023 \leq n \leq 0.027$ for smaller materials. Equation [4] yields $0.036 \leq n \leq 0.042$ for boulder gravels and $0.029 \leq n \leq 0.035$ for the rest. The second set of values seems to be rather high.

Velocity may now be calculated from the familiar Manning formula. Using Strickler's n, values in excess of 6 m s^{-1} are found, which are improbably high. For Limerinos' n, 1.6 m s$^{-1} \leq v \leq$ 4.8 m s^{-1}. Comparison with velocity criteria for designed cobble–gravel channels (Fig. 7) suggests that the Limerinos results are best for all but the apparently shallowest channels. The Strickler approach yields velocity estimates that are least sensitive to errors in the input parameters (see Appendix 2), which will be of greatest proportional importance in the shallowest flows.

An interesting pattern emerges from this work: the channel bottom boulder gravels (except T1) yield much greater depths than do the other deposits. In particular, the sequence of T2 sediments returns results consistent with sampling position. There is no doubt that reconstruction is sensitive to material sampling site, a factor that has not heretofore been emphasized in palaeohydraulic reconstructions. It presents serious constraints for studies from limited outcrop.

An alternative interpretation may be placed upon the results from T2 and T3 fine gravels. If they are backwater deposits, as has been proposed, then the slope of the terrace surface, used in the calculation, may be a large overestimate of the water surface gradient at the time. That gradient remains unknown. The ratios of boulder gravel D_{90} grain sizes to fine gravel D_{90} grain sizes in T2 and T3 yield values of 6 and 4, respectively. This indicates the reduction in gradient to move the materials composing the fine gravels, with no other hydraulic changes. One would expect a substantial increase in depth in backwater as well, which would further reduce the requisite gradient. Adopting $S = 0.5 \times 10^{-3}$ as a figure of merit for both cases

TABLE 2. Data of palaeohydraulic computations

Datum	T1		T2			T3		Contemporary river	
	bg[a]	cg	bg	cg	fg	bg	fg	Bridge	Orchard bar
Measured									
S	0.0026		0.0036			0.0040		0.0017	
D_{b50} (mm)	51	33	135	66	16	148	23	44[b]	70
D_{b84} (mm)	158	79	267	123	41	345	72	179	365
D_{b90} (mm)	180	88	283	139	45	374	90	195	390
D_{s50} (mm)								109	175
Computed									
D_{b90}/D_{b50}	3.5	2.7	2.1	2.1	2.8	2.5	3.9	4.4[b]	5.6
τ^*_{ci}	0.028	0.035	0.044	0.043	0.034	0.037	0.025	0.020	0.020
τ_{ci} (kg m^{-2})	8.32	5.08	20.55	9.86	2.53	22.83	3.71	6.44	12.87
d (m)	3.2	2.0	5.7	2.7	0.7	5.7	0.9	3.8	7.6
n: Strickler	0.029	0.025	0.031	0.027	0.023	0.032	0.025	0.029	0.032
n: Limerinos	0.036	0.032	0.040	0.035	0.029	0.042	0.033	0.037	0.042
v: Strickler (m s^{-1})	3.8	3.2	6.2	4.4	2.1	6.3	2.4	3.5	5.0
v: Limerinos (m s^{-1})	3.1	2.5	4.8	3.4	1.6	4.8	1.8	2.7	3.8
q: Limerinos (m^3 s^{-1} m^{-1})	9.9	4.9	27.4	9.2	1.1	27.5	1.7	10.3	28.9

[a]Material descriptors are as in Fig. 6.
[b]See Appendix 1.

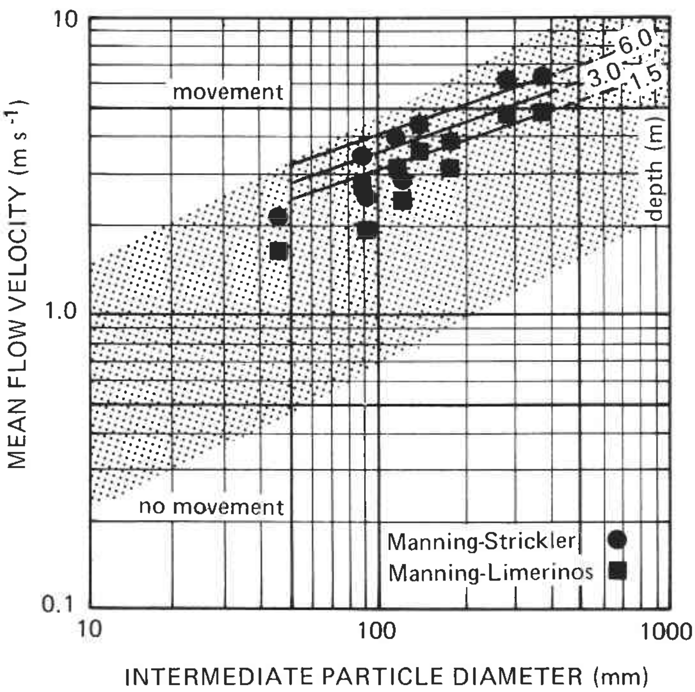

FIG. 7. Comparison of computed velocities and competence curves of Neill (1973). The stippled area covers the region of mean velocity and grain-size observations in boulder gravel stream channels.

(which is an eightfold reduction) yields the results of Table 3. They compare very well with channel gravel figures in Table 2. The calculation is arbitrary but confirms that backwater deposition of the extensive fine gravel deposits is a reasonable proposal, since q is now more nearly comparable with the values computed for the boulder gravels.

To make further progress, we require knowledge of channel geometry. Under T2, T3, and T4 on the left bank of the Fraser River, channel sections are exposed (Figs. 4*b*, 5, 6). Table 4 gives the measured geometry for these channels. The T2 section is complete but seems small. The section under T3 is triangular. The right bank is missing, having been truncated by subsequent degradation. However, the extant morphology makes clear that the right channel margin was the slip-off slope of a leftward-migrating channel. The sloping channel bed has been extended with some confidence to meet the projected level of overbank deposits. Data of the reconstructed T3 channel are comparable with the dimensions of the present channel. The channel under T4 is ellipsoidal in section. Its right side also is missing: there is no obvious basis for reconstruction.

Comparison of the reconstructed channels with palaeohydraulic depths reveals rough agreement. The variations are not surprising, since the sediment samples were taken at specific sites that may not be representative of the entire channel. In the modern channel, for example, a bar head site was selected near a bridge pier that revealed a high-water line about 3 m above the surface (reached in about 4 years since construction on the bridge), in comparison with a 6.5 m mean depth for the entire channel. It should be apparent as well (see Appendices 1, 2) that the computation of τ^*_{ci} leaves sufficient leeway for the observed discrepancies between computed and geometric depths.

To complete discharge calculations, the Manning–Limerinos n computed for the boulder gravels (Table 2) is adopted and used with *observed* channel geometry to estimate the velocity for the entire channel, thence the discharge.

Palaeoflow and comparison with the modern river

From data of Tables 2 and 4 the results of Table 5 were obtained. For the modern river, n was selected from the "Orchard bar" site, where the most typical bed was examined, but the channel dimensions were derived from the drawings for the new bridge. The river is gauged 13 km downstream near Texas Creek. Figure 8 is the annual flood frequency graph adjusted for Seton River outflows. Table 5 includes hydraulic data for the gauging station and adjusted data for the river at Lillooet that are based upon observed flows. The calculated flow for the river at Lillooet falls midway between the mean

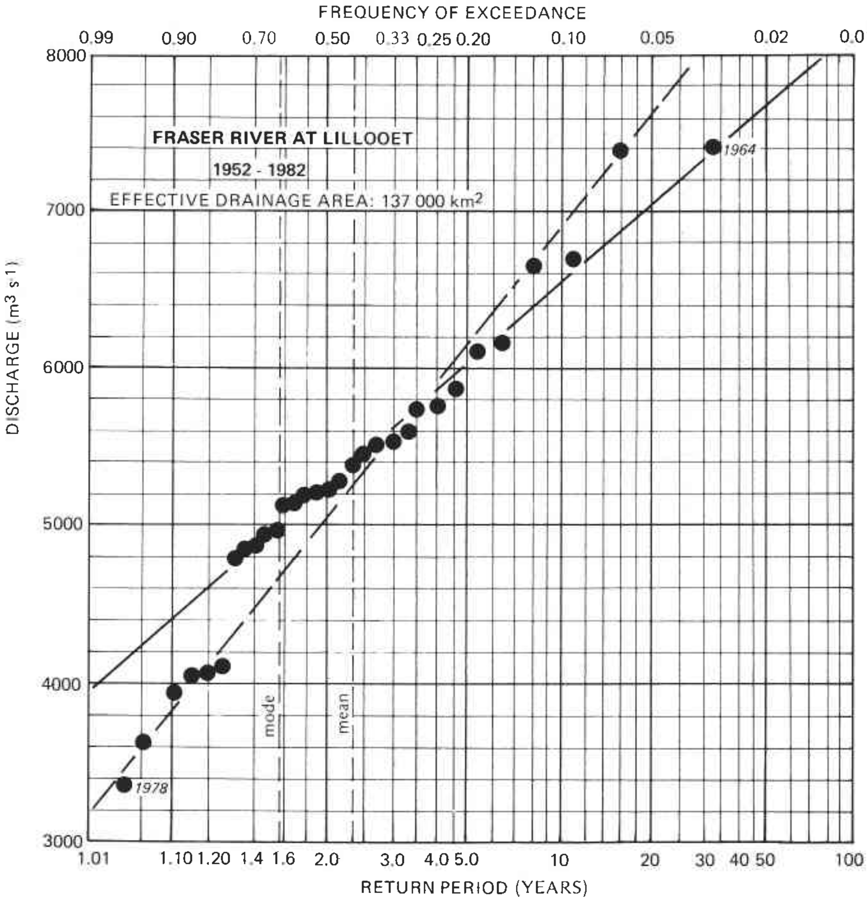

FIG. 8. Return period plot for annual maximum daily flow in the Fraser River at Lillooet. Data adjusted from Water Survey of Canada (WSC) Station 08MF040 (Fraser River near Texas Creek) by subtraction of measured flows in the Seton River. The two lines represent two interpolations of an underlying flood sequence.

TABLE 3. Palaeohydraulic calculations for fine gravels on T2 and T3 based on $S = 0.0005$

Datum	T2	T3
d (m)	5.0	7.4
n^a	0.028	0.030
v (m s^{-1})	2.3	2.6
q (m^3 s^{-1} m^{-1})	11.7	19.2

[a]Based on [4].

TABLE 4. Data of exposed palaeochannels of the Fraser River

Terrace	Width on outcrop	w_s^a (m)	d_{max} (m)	A^b (m^2)	d_*^c (m)
T2		96	8.9	330	3.4
T3	260[d]	157	16.0	1250	8.0
T4[e]	165	121	12.3	990	8.2
Modern channel[f]		275	13.9	1775	6.5

[a]Exposed width · sin a, where a is the angle between the outcrop and the palaeochannel direction seen on the terrace surface. For T2, true width was measured on the surface.
[b]Adjusted as width.
[c]$= A/w_s$, the hydraulic mean depth.
[d]Extrapolated from partial section. See text.
[e]Incomplete. See text.
[f]From British Columbia Ministry of Transport and Communications bridge engineering geotechnical report.

annual flood and the 10 year flood. This establishes some confidence in the flow reconstruction procedure. It is also salutary to note that at the bridge site, the estimated flow depth to move the materials present there, 3.8 m (Table 2), is not much greater than the 3 m waterline observed on the bridge. (In Appendix 1 we speculate that this material may not have been disturbed since the bridge was built.)

The T3 flow is slightly in excess of the modern 30 year flood and is about 1.4 times the contemporary mean annual flood. This number falls within the range of proportional errors enumerated in Appendix 2. Nonetheless, it is consistent with the large boulder sizes and high gradient to suggest larger flood flows then. The result falls just outside the range of changes in flood flows observed on British Columbia rivers in the twentieth century (Barrett 1980) and thus is certainly reasonable.

The T2 palaeochannel yields a result only 0.2 times the modern mean annual flood. This result falls outside the reasonable range of errors for the calculations. We cannot escape the conclusion that flows on T2 occupied multiple channels.

Some further insight into this appearance might be gained from a morphological comparison. From a recent reassessment of data of the single thread – braided transition in gravel bed channels, it appears that a useful discrimination for British Columbia cobble and boulder gravel rivers is

$$[5] \quad S\sqrt{Q} \leq 0.07$$

TABLE 5. Palaeoflow and contemporary flow estimates of the Fraser River near Lillooet

Station	Q_{30} ($m^3 s^{-1}$)	Q_{10} ($m^3 s^{-1}$)	$Q_{2.3}$ ($m^3 s^{-1}$)	Q_{est} ($m^3 s^{-1}$)	w_s[a] (m)	A (m^2)	d_* (m)	v ($m s^{-1}$)	n	S
T2				1100	96	330	3.4	3.4	0.040	0.0036
T3				7500	157	1250	8.0	6.0	0.042	0.0040
Fraser R. at Lillooet (u/s Seton R.)										
From morphology				6035	275	1775	6.5	3.8	0.042	0.0017
From flows	7300	6550	5400		275	1775	6.5	3.1	0.037	0.0017
Fraser R. near Texas Cr. (08MF040)	7500	6800	5430		190	1540	8.1	3.5	0.035[b]	0.00093[c]

[a]All the following columns pertain to the estimated flow or to the $Q_{2.3}$ (mean annual flood) flow for the modern measurements.
[b]n assigned, then v, A, and d_* calculated for this station only, using measured w_s.
[c]From large-scale topographic map.

(see Kellerhals (1982) for a discussion of the data). For $Q = 5400$ m³ s⁻¹, $S = 0.95 \times 10^{-3}$. The contemporary river lies just above this limit, and indeed, this corresponds to the appearance of the river, which exhibits a few emergent bars to very high flows in the Lillooet reach. In T3 time, the gradient appears to have exceeded the limit about fourfold. However, the channel was being cut into Lillooet silt: it is likely that the confining silt bluffs (which may likewise affect the river today) prevented the development of a braided channel. The upper terraces, with a gradient two to four times larger than the threshold and no apparent constraint, almost certainly carried braided channels. Even if we were to suppose T1 – T2 flows as small as one half of recent flows, the threshold for braiding would rise to only 1.3×10^{-3}. The braided channel in T1 – T2 time renders the calculation of a complete palaeohydrology impossible.

Conclusions

A suite of Holocene fluvial terraces occurs along the Fraser River at Lillooet, where the Fraser Valley widens uncharacteristically to several kilometres breadth. The terrace development here is exceptional for many kilometres along the river, probably because of the room afforded by the physiographic setting. Of the eight major surfaces, four on either bank, the upper four are nonpaired and the lower four form two pairs.

A broadly consistent alluvial stratigraphy underlies the terraces, comprising overbank fines, stratified channel gravels, and boulder gravels in thin beds interpreted as channel bed deposits. Aeolian sediments cap the terraces and obscure the former riverine channel topography. Glaciofluvial sediments and glaciolacustrine Lillooet silt underlie the alluvium.

The presence of boulder gravel at several levels within some sections, the paired lower terraces, the vertically stacked channels (at T4 level), and the extensive fine gravel beds all suggest that episodes of aggradation interrupted the general degradation of the Fraser River. The fine gravels suggest deposition in backwater under downstream control.

At Jones bench and Texas Creek, 6 and 17 km downstream from Lillooet, respectively, postglacial landslides may have held up the river for many years. They represent a plausible mechanism for backwater development. Palaeohydraulic calculations indicate that this explanation is consistent with sediment of the observed calibre having been deposited on the bed of a river of the magnitude of the Fraser River.

A channel section preserved in the T3 terrace indicates that in T3 time flood flows may have exceeded contemporary floods. Leaving aside computational errors, the conclusion remains tentative, since the recurrence interval of bankfull flow is not known for that time. A T2 channel section carried much smaller flows, confirming morphological indications of braiding.

Chronological control of this history is poor. The presence of Mazama tephra under the aeolian capping on the T1 surface and its apparent absence everywhere else suggest that the river may have remained little incised or unincised throughout most of Hypsithermal time. The substantial extent of this surface lends some further support to this notion, although of course the river would have become more confined laterally with degradation in view of the increasing volumes of bank material to be eroded. Charcoal dated at about 1200 ^{14}C years BP, which is younger than overbank sediments and overlain by fine channel gravels at site 15 in L4, provides a maximum age for a relatively recent phase of aggradation and related terrace formation by the Fraser River. This relatively young age for a terrace that is now 28 m above river level and the implied rate of subsequent downcutting are in accordance with the process rates that would be necessary for terrace formation in Lillooet basin to be a post-Hypsithermal phenomenon.

The role of the palaeohydraulic calculations in this study is instructive. Despite reasonable exposure on the east bank, the variability of the alluvial gravels remains so great and the critical evidence of channel geometry (including gradient) remains sufficiently obscure that only approximate results (one is inclined to say ‘‘order of magnitude results’’) may be achieved. Such calculations are useful only in juxtaposition to well-defined morphology and stratigraphy and are best used to support the interpretations derived therefrom. This means that palaeohydraulic exercises remain essentially qualitative.

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Appendix 1: Texture of the alluvial sediments

Determination of a "representative" grain size for hydraulic computations is a complex matter that has been given insufficient attention. It may be divided into (*i*) choice of sampling site and (*ii*) sampling procedure.

Given the capacity of rivers to sort sediment into locally homogeneous deposits that vary remarkably from place to place in the river bed, one doubts that any single sample or small number of samples could be representative for the river. At best, a sample will represent the local conditions of deposition. This appears to constrain the possibility of computing palaeohydraulics for an entire river. Since calculations are based on the force balance at the boundary between the drag exerted by the flow and the inertia of the particles, the most likely site to obtain data descriptive of the main flow is the thalweg. This is inaccessible (except with very elaborate arrangements) in a river as large as the Fraser River. Bar head positions have been used as a surrogate site, since substantial flow impinges there.

In deposits exposed as outcrop, it may be possible to sample the main channel bed. In this study, we attempted to do that. For comparison, we also sampled higher positions in a section occupied by stratified "bar gravels." What is impossible to determine, given the restricted accessible outcrop, is the variability of material within the sedimentary unit. In fact, logistical constraints restricted our examination to one sample per sampled bed. We have some reservations about the representativeness of the samples in Table A1. In T1 and T2 we chose the only well-exposed sites. Additional sites are available in T3, but they would be hazardous to examine.

Sampling procedures depend upon the necessary size of a sample for adequate characterization of a deposit. That depends upon the size of the largest clasts present, since there must be sufficient material contained in the sample to permit the proportion in the largest size class to be assessed accurately. There does not exist a well-written standard in this matter. (A recently promulgated International Standards Organization (ISO) standard is defined in terms of the statistics of the sample, which are not known at the time of sampling.) We have in recent work attempted to ensure that the largest clast always constitutes less than 1% of the sample weight, and we have gained some analytical confidence in this procedure.

In sampling Fraser River boulder gravels, one encounters clasts of the order of tens of kilograms in weight, which would require a sample of some tonnes to meet the above criterion. Using hand methods, we were constrained to accept samples in the range 200–600 kg. Data of Table A1 show that the 1% criterion was not met. Since the proportional distribution depends upon all material, the texture analyses must be regarded as possibly biased, probably toward the smaller sizes. However, it is likely that the largest size class present has been identified. The result is that the D_{90} data are probably reasonable and the D_{50} data are perhaps too small.

Sampling followed these procedures: (1) Shovel material from the outcrop into buckets and weigh (using a 20 kg spring balance). (2) Sieve in the field down to -4.5ϕ by 0.5ϕ intervals, weigh the size classes, and discard the material. (3)

TABLE A1. Grain-size sample characteristics

Sample	Site	Sample Weight (kg)	Proportion: Boulder	Cobble	Pebble	Sand	Largest clast: b axis (mm)	Weight (kg)	% by wt. of sample	Remarks
1	Contemporary river: bar head at new bridge (u/s Seton R. mouth)	190	—	0.23	0.44	0.33	215	11.5	6.1	Very high sand percent: sample atypical of fluvial gravel
2	T2 Riverlands exposure: sandy, pebbly gravel near top of section	1.61	—	—	0.55	0.45	16	0.006	0.4	
3	T2 stratified channel gravel: high in section	20.4	—	0.03	0.76	0.21	90	0.63	3.1	Bag sample: no field analysis
4	T1 Hwy 12 exposure: stratified channel gravel high in section	233	—	0.27	0.49	0.24	146	3.4	1.5	Clast-supported deposit
5	T1 Hwy 12 exposure: boulder gravel from base of unit	305	0.05	0.39	0.42	0.14	305	16.5	5.4	Matrix-supported boulders
6	T3 bridge exposure: boulder bed; position in unit uncertain	485	0.30	0.45	0.13	0.12	400	73.1[a]	15.1	Boulder framework deposit: remarkably biomodal
7	Contemporary river: left bank channel bar d/s Seton R. (Orchard bar)	574	0.21	0.30	0.38	0.11	447	93.0[a]	16.1	Second stone (264 mm) is 4.9% of sample: +800 mm material is present on bar
8	T2 palaeochannel exposure: stratified channel gravel high in section	224	—	0.51	0.35	0.14	181	6.9	3.1	Framework supported
9	T2 palaeochannel exposure: boulder gravel from base of channel	376	0.21	0.45	0.26	0.08	310	52.2[1]	13.9	Framework supported
10	T3 near bridge: pebbly gravel from top of bed	151	—	0.19	0.63	0.18	128	2.6	1.7	
BCMTH	Subsurface sample at new bridge: −2 m	?	—	0.18	0.61	0.21	100	?		From test pit in bar for bridge foundation: probably a large sample

[a]Clasts to 21 kg are weighed directly. Larger clasts are estimated as $\rho_s V = 2.77abc/6$, where a, b, and c are the principal axes. The formula has been extensively confirmed by comparisons.

Split finer material on a quartering cloth to a 6−8 kg subsample for return to the laboratory.

Laboratory procedures were standard. It was assumed that moisture (always of the order of 1% or less) was entirely associated with the fine fraction. The sand fraction sieved was typically of the order of 0.05% of the sample and 0.25% of the sand portion of the sample. However, analyses in this range have been confirmed to be accurate.

An additional problem is that on a vertical outcrop, it is difficult not to lose some material during sampling. It is probable that mainly fines are lost. No attempt has been made to assess this effect.

The quality of the samples may be assessed further by considering the grain-size distribution. For normal fluvial gravels we expect good sorting, so that D_{90} is a small multiple of D_{50}. This is broadly realized in the samples, with $D_{90} \leq 4D_{50}$ except in the contemporary river. We further expect $D_{s50} \sim 2.5D_{b50}$ (Parker *et al.* 1982); at the bridge, $D_{s50} = 13.4D_{b50}$. Again, the proportion of material smaller than -1ϕ should be less than about 0.25, which is the limit set by packing and porosity for fines that have infiltrated a framework-supported gravel. The bridge site has more than 0.33 finer than -1ϕ. Finally, we expect the upper limits of surface and subsurface samples to converge. This occurs for the samples for which there are surface counts, except that at the bridge.

It is concluded that the bridge subsurface samples are atypical. They have possibly been influenced by changes in sedimentation since construction of the new bridge (1979−1980) or possibly still reflect construction disturbance. Texture parameters at the downstream Fraser River bar site ("Orchard bar") are consistent and seem to be much more typical for the Fraser River. Texture parameters for the contemporary river are taken from the Orchard bar site. The bridge samples, taken in the section for which complete channel geometry is available, are retained for a check calculation.

Appendix 2: Precision of the calculations

The precision of the palaeohydraulic calculations is limited by the precision of the input data. The effect of a known error measurement can be computed through the simple power equations used in the calculations.

Equation [2] is used to set the nondimensional shear to move a grain of given size in a mixture of grains according to the ratio $R = D_i/D_{b50}$, the argument of the equation. We obtain

$$d\bar{\tau}^*_{ci}/dR = 0.0834(-0.872)R^{-1.872}$$

Dividing left and right sides by the corresponding expressions in [2] and transposing dR yields

$$d\bar{\tau}^*_{ci}/\bar{\tau}^*_{ci} = -0.872dR/R$$

i.e., the proportional change in $\bar{\tau}^*_{ci}$ is −0.872 times the propor-

TABLE A2. Proportional error, $\triangle n/n$ in the estimation of the resistance coefficient, and $\triangle v/v$ in the estimation of velocity[a]

Basis for n	$\partial\bar{\tau}^*_{ci}/\bar{\tau}^*_{ci}$	$\partial D/D$	$\partial S/S$	$\partial D_{84}/D_{84}$	$\partial n/n$
Strickler		0.17			
(eq. [3])	0.67	0.50	−0.17		
Limerinos[b]	−0.51	−0.51	0.60	0.71	
(eq. [4])	1.06	1.06	−0.56	−0.56	
n arbitrarily					
assigned	0.67	0.67	−0.17		−1.0

[a]The first of the two entries under each heading is for the resistance coefficient, the second for velocity. No entry indicates no effect.

[b]Results are only approximate, as they entail an arbitrary evaluation of the parameter X (see text).

tional change in R.

We have discussed in Appendix 1 possible sources of error in grain-size specification and considered that D_{b50} may be negatively biased (by up to 5.4 times in the extreme case of the bridge site. That would yield underestimates of $\bar{\tau}^*_{ci}$ of up to 4.7 times).

Further, [2] is statistical, with ±22.3% variation in the α = 0.05 range of estimates of $\bar{\tau}^*_{ci}$ (Andrews 1983). Since there is no basis, short of additional sampling, to assess the actual bias in D_{b50} values, we will consider only the $0.22\bar{\tau}^*_{ci}$ statistical error in what follows.

Combining [1] with $d = \tau_{ci}/\gamma S$ yields an expression for flow depth that is linear in each of $\bar{\tau}^*_{ci}$, D_i, and S. We obtain

$$\partial d/d = \partial D_i/D_i; \qquad \partial d/d = \partial\bar{\tau}^*_{ci}/\bar{\tau}^*_{ci}; \qquad \partial d/d = -\partial S/S$$

Insofar as the choice of D_i is arbitrary, its error is immeasurable. We have judged that D_{90} probably is correctly determined to ϕ class for the site at which the texture of the sediment was measured (though not necessarily as an integral measure for the channel), whence it is probably wrong by no more than ±0.19 times. Errors in gradient (Table 1) are apt to be less than ±0.2 times for terraces on which palaeohydraulic computations have been essayed. Combining these sources of error yields, for depth of flow, a range of 0.52−1.7 times the estimate.

The Strickler estimate of Manning's n ([3]) depends only upon grain size in the 1/6 power. Hence, $\mathrm{d}n/n = (1/6)\mathrm{d}D_i/D_i$.

Equation [4], Limerinos' estimate of n, leads to more complicated derivatives. Let $Z = \bar{\tau}^*_{ci}s_s' D_i/\gamma S$, which is linear in each of the variates $\bar{\tau}^*_{ci}$, D_i, and S; and let

$$X = 1.16 + 2.0 \log (Z/D_{84})$$

Then, rewrite [4] as

$$n = 0.113Z^{1/6}/X$$

Then we obtain the following results by partial differentiation:

$$\partial n/n = [(1/6 - 2.0)/X]\ \partial\bar{\tau}^*_{ci}/\tau^*_{ci}$$
$$\partial n/n = [(1/6 - 2.0)/X]\ \partial D_i/D_i$$
$$\partial n/n = [(-1/6 + 2.0)/X]\partial S/S$$
$$\partial n/n = (2.0/X)\partial D_{84}/D_{84}$$

Noting that $\bar{\tau}^*_{ci} \sim 10S$, in general, and $D_i/D_{84} \sim 1$ (we choose $D_i = D_{90}$), then $Z/D_{84} \sim 16.5$ and $X \sim 3.6$. The expressions given above then lead to the results tabulated in Table A2, which are not exact because of the approximations just made.

The analysis for velocity is similar and yields

$$\partial v/v = (1/2 + 2/X)\ \partial\bar{\tau}^*_{ci}/\bar{\tau}^*_{ci}$$
$$\partial v/v = (1/2 + 2/X)\partial D_i/D_i$$
$$\partial v/v = (-2/X)\partial S/S$$
$$\partial v/v = (-2/X)\partial D_{84}/D_{84}.$$

Then the outside error of an estimate of n or v is

$$\prod_{i=1}^{4} [(1.0 \pm c_i)\ (\partial p_i/p_i)]$$

where the c_i are coefficients from Table A2, and the p_i represent each contributing parameter in turn. For Limerinos' n, this yields limits 0.63−1.52 times the estimate.