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Modeling of Paleohydrologic Change during Deglaciation
Modélisation des changements paléohydrologiques survenus au cours de la déglaciation
Modellierung der paleohydrologischen Veränderungen während der Enteisung

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Résumé de l'article

Le présent article vise à vérifier si les estimations de paléoécoulement faites sur les séquences de terrasses proglaciaires (à partir des réseaux de chenaux, des changements morphologiques et sédimentaires) confirment le modèle des changements à grande échelle qui ont affecté le régime hydrologique au cours de la déglaciation. La modélisation de l'écoulement d'anciens cours d'eau anastomosés est toutefois encore au stade préliminaire en raison des nombreuses sources d'erreur et des hypothèses à vérifier. II faut tenir compte sur le terrain de la largeur, de la profondeur, de la pente et de la nature sédimentaire des chenaux; de la modélisation théorique de la profondeur du débit, des coefficients de résistance, de la vitesse, des nombres de Froude et de l'écoulement ; de l'interprétation géomorphologique des écoulements en voie de modélisation. Les divers modèles d'écoulement ont fourni des estimations approximatives sur les anciens débits maximaux et sur leurs variations au cours de la déglaciation. Celle-ci, étalée sur une longue période, semble avoir été associée à de fortes diminutions des débits maximaux (de 10 à 30 fois moindres). Sur de plus courtes périodes, les débits maximaux provoqués par les eaux de fonte et par des phénomènes comme les inondations associées au volcanisme masquent les tendances de l'écoulement à plus long terme.

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MODELING OF PALEOHYDROLOGIC CHANGE DURING DEGLACIATION*

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ABSTRACT This paper aims to test whether estimates of paleodischarge made for proglacial terrace sequences support the model of largescale change in hydrologic regime (as inferred from channel pattern, form and sedimentology changes), during deglaciation. Modeling paleodischarges of former braided stream systems on each terrace surface is shown to be at a tentative stage, however, because of numerous error sources and assumptions that remain to be tested. These are discussed in relation to field measurements of channel width, depth, gradient and sedimentology; to theoretical modeling of flow depth, resistance coefficients, velocity, Froude number and discharge; and to geomorphic interpretation of the flow events being modeled. The different discharge models adopted have provided approximate estimates of former peak flows and patterns of change in peak flow magnitudes during deglaciation. Longterm deglaciation appears to have been associated with largescale decreases in peak flows of between 10 and 30 times. On shorter timescales, peak flows produced by shortterm meltwater and by non-melt processes, such as volcanically induced floods, mask any longer term discharge trends in the catchment.

RÉSUMÉ *Modélisation des changements paléohydrologiques survenus au cours de la déglaciation.* Le présent article vise à vérifier si les estimations de paléoécoulement faites sur les séquences de terrasses proglaciaires (à partir des réseaux de chenaux, des changements morphologiques et sédimentaires) confirment le modèle des changements à grande échelle qui ont affecté le régime hydrologique au cours de la déglaciation. La modélisation de l'écoulement d'anciens cours d'eau anastomosés est toutefois encore au stade préliminaire en raison des nombreuses sources d'erreur et des hypothèses à vérifier. Il faut tenir compte sur le terrain de la largeur, de la profondeur, de la pente et de la nature sédimentaire des chenaux; de la modélisation théorique de la profondeur du débit, des coefficients de résistance, de la vitesse, des nombres de Froude et de l'écoulement; de l'interprétation géomorphologique des écoulements en voie de modélisation. Les divers modèles d'écoulement ont fourni des estimations approximatives sur les anciens débits maximaux et sur leurs variations au cours de la déglaciation. Celle-ci, étalée sur une longue période, semble avoir été associée à de fortes diminutions des débits maximaux (de 10 à 30 fois moindres). Sur de plus courtes périodes, les débits maximaux provoqués par les eaux de fonte et par des phénomènes comme les inondations associées au volcanisme masquent les tendances de l'écoulement à plus long terme.

ZUSAMMENFASSUNG *Modellierung der paleohydrologischen Veränderungen während der Enteisung.* Dieser Artikel will überprüfen, ob die Schätzungen der Paleowasserführungen, die für die proglazialen Terrassensequenzen gemacht wurden, das Modell eines tiefgreifenden Wechsels im hydrologischen System während der Enteisung stützen (so wie es aus dem Netz der Kanäle und den morphologischen und sedimentologischen Veränderungen abgeleitet werden kann). Die Modellierung der Paleowasserführungen von zuvor verwilderten Wasserlaufsystemen auf jeder Terrassenoberfläche befindet sich jedoch noch in einem Versuchsstadium, da zahlreiche Fehlerquellen und Annahmen noch erst überprüft werden müssen. Diese werden in Bezug auf Feldmessungen der Kanalweite, Tiefe, der Neigung und der Sedimentologie diskutiert; ferner in Bezug auf die theoretische Modellierung der Wasserflußmenge, die Widerstands-koeffizienten, die Geschwindigkeit, die Froud-Zahl und den Wasserertrag; und in Bezug auf die geomorphologische Interpretation der modellierten Fließvorgänge. Die verschiedenen angenommenen Wasserflußmodelle führen zu ungefähren Schätzungen früherer maximaler Fließvorgänge und Modellen ihrer Veränderungen während der Enteisung. Über einen langen Zeitraum scheint die Enteisung mit einer starken Abnahme der maximalen Fließwerte um das 10 bis 30 fache verbunden gewesen zu sein. Über Kürzere Zeitspannen verdecken maximale Fließwerte, die durch Kurzzeit-Schmelzwasser und andere vom Schmelzen unabhängige Prozesse wie z.B. durch Vulkane hervorgerufene Überflutungen produziert werden, die Fließtendenzen im Einzugsgebiet über einen längeren Zeitraum.

* Contribution du premier symposium de la CANQUA, sous la direction de René W. Barendregt

INTRODUCTION

Field evidence from a number of mid-latitude areas that were subject to deglaciation at the end of the Late Glacial or Late Wisconsinan indicates that this period was associated with major changes in river channel pattern, form, sedimentology and stability (*e.g.* KOZARSKI and ROTNICKI, 1977; MYCIELSKA-DOWGIALLO, 1977; KOUTANIEMI, 1980; KOZARSKI and TOBOLSKI, 1981; STARKEL, 1981, 1983; SZUMANSKI, 1981; KOZARSKI, 1983; MAIZELS, 1983a, b, c). These changes reflected the exceedance of the thresholds for channel change, such that steeply graded, low sinuosity, multi-thread bedload channels characteristic of glacial meltwater flow regimes, became replaced by more gently graded, high sinuosity, single-thread suspended load channels that now characterize many non-glacial mid-latitude river systems (Fig. 1). Also, the exceedance of the stability threshold is extensively reflected in the periodic episodes of downcutting into the outwash valley fill deposits to form distinctive terrace sequences, many of which exhibit extensive braided paleochannel systems.

Although these channel changes have been interpreted as reflecting largescale decreases in discharges at the end of the period of glaciation, no systematic attempt at establishing the discharge changes associated with deglaciation terrace sequences and related paleochannel systems has been made. The object of this paper is to test whether estimates of paleodischarge made for proglacial terrace sequences support the model of largescale change in hydrologic regime (as inferred from channel pattern and form changes) during deglaciation.

The paper examines some of the characteristics of meltwater discharge variations during deglaciation, and how different flow events may be represented in the terrace sequence. The paper then outlines the field and theoretical procedures adopted for estimating paleodischarges, together with some discussion of the likely sources of error. Finally, the paper applies the procedures outlined to a series of field sites, representing proglacial terrace sequences that formed during deglaciation over different timescales.

PROPOSED MODEL OF PALEODISCHARGE CHANGES DURING DEGLACIATION

While the field evidence indicates largescale changes in channel pattern, form, sedimentology and stability, at the end of a glacial phase, three major problems arise in the determination of associated changes in paleodischarge. First, the procedures presently available for determining the discharges of former braided stream environments rely on a still rather crude and inaccurate methodology. Second, the nature of the discharge event(s) being predicted, and their magnitude and frequency relative to the complete flow record, often remains indeterminate. Since the nature of the flow event may remain uncertain, a third problem arises in the interpretation of long term paleodischarge changes. Uncertainties remain in determining the extent to which the estimated paleodischarge changes reflect real overall changes in discharge in the catchment, rather than a sequence of random, episodic events

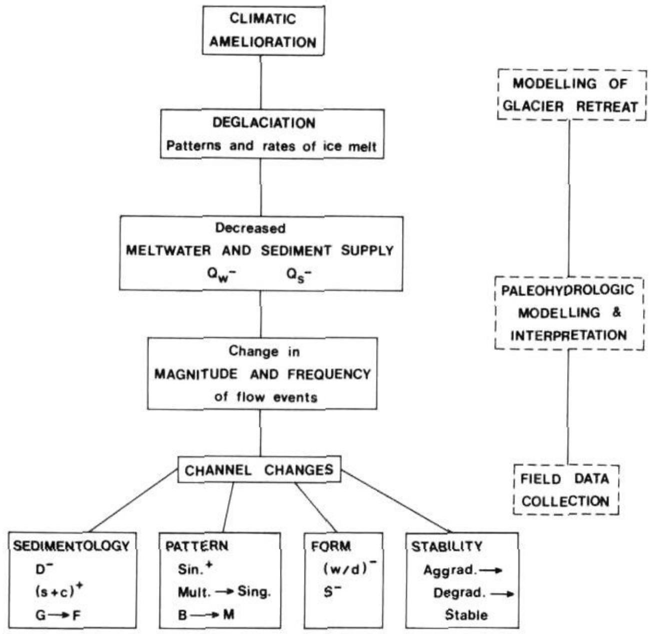

FIGURE 1. Basic assumptions of paleohydrologic analysis of proglacial terrace deposits.

Hypothèses de base de l'analyse paléohydrologique des dépôts proglaciaires de terrasses.

unrelated to longterm catchment changes. The latter two problems are discussed below, while the discharge prediction procedures are outlined later in the paper.

NATURE OF THE FLOW EVENTS REPRESENTED IN THE TERRACE SEQUENCE

Proglacial meltwater regimes are subject to temporal fluctuations on a wide range of scales, from instantaneous pulses, through diurnal flood events, seasonal peaks, and major episodic 'glacier burst' or 'jökulhlaup' floods. Cyclic flow events are largely ablation-controlled, but many peak flows may be produced by other mechanisms such as precipitation storm events, sudden drainage of ice-dammed lakes, and subglacial volcanic eruptions. Hence the proglacial channel system may have been formed by flows produced by any of these events. The possible composite origin of the discharges which helped create the proglacial channel system makes it difficult to assess the true nature of the flow event(s) that are represented in each terrace surface channel system. This in turn makes questionable the direct comparison of paleodischarges for different channel systems on successive terrace surfaces both in a single area and in different areas.

An attempt has been made in this paper to overcome this dilemma by sampling the coarsest channel sediments present, since they are likely to represent the peak flows in each of the channel systems being investigated. This approach increases the likelihood that the flow events represented by the channel systems on successive terrace surfaces are more directly comparable in terms of the relative magnitude-frequency relations that they represent. However, the origin of the peak flows recorded, and the relative magnitude-frequency conditions that they do represent, remain uncertain.

A further assumption of this approach is that the channel systems being investigated were sufficiently sensitive to changing peak flows that the latter were actually reflected in the channel systems themselves. Hence, the observed alluvial channel pattern, form and sedimentology are assumed to be directly related to the last series of peak flows which occupied each successive channel system prior to its abandonment and subsequent incision to form a terrace surface. This interpretation implies, however, that terrace incision occurred during even higher magnitude flows (and/or during a relative decrease in sediment supply) than those represented by the terrace surface channels. This in turn suggests that the paleoflow conditions estimated for the braided channel systems may well represent the minimum flow conditions required for local terrace incision, and hence may represent local thresholds for largescale channel change.

LONGTERM CHANGES IN PEAK DISCHARGE DURING DEGLACIATION

The uncertainties about the nature of the flow events represented in a terrace sequence compound the difficulties involved in determining the pattern and scale of discharge changes that were associated with deglaciation. The variations in peak discharges recorded on successive terrace surfaces may or may not reflect significant temporal changes in peak meltwater flows from a declining ice mass. The change magnitudes of former peak flows may or may not reflect flows of equivalent frequency through the period of deglaciation. Hence, the assumption that peak meltwater flows will decrease at the end of a glacial phase is likely to be acceptable only in areas of largescale and/or longterm deglaciation where an ice mass has almost completely disappeared from the catchment. Where shorter term ice retreat or glacier fluctuation occur, any significant peak discharge changes that reflect simple changes in ice mass are likely to be masked by the superimposition of flow fluctuations of different periodicities.

Nevertheless, threshold ice mass conditions, or threshold rates of ice melt, may exist for which decreases in discharge become significant in terms of channel change once deglaciation has reached a critical stage. This paper examines the extent to which deglaciation over different temporal and spatial scales affects the degree of discharge change predicted through a deglaciation terrace sequence.

SELECTION OF FIELD SITES

The paleohydraulic modeling procedures discussed below have been applied to 6 different deglaciation terrace sequences that have experienced deglaciation over a range of different timescales. These areas have been selected because each exhibits a distinct terrace sequence apparently formed in association with major hydrological changes that occurred during deglaciation.

The areas selected have experienced either a single major phase of deglaciation (North Esk, West Greenland, and Norway), or multiple phases of glaciation and deglaciation, or glacial fluctuation (Tekapo, N.Z. and Iceland). The timescales involved were selected to range from (a) Late-Glacial deglaciation in which ice has disappeared totally (N. Esk), partially (Tekapo) or only marginally (West Greenland) from the catchment, (b) "Neoglacial" deglaciation (S. W. Greenland, Iceland), and (c) Little Ice Age deglaciation, with ice retreating significantly in the catchment (Iceland and Norway). The areas have also been selected to include both known non-jokulhlaup and jokulhlaup glaciers (Iceland), in order to gauge the influence of largescale episodic flood events on longterm hydrological trends associated with deglaciation.

FIELD PROCEDURES FOR PALEODISCHARGE PREDICTION

The determination of the main paleohydraulic parameters required for discharge prediction relies on a large number of assumptions, each of which may give rise to errors that reduce the reliability and accuracy of the computed estimates (see Table I). Although reference is made to these different error sources, detailed assessment of their significance is not possible in this summary paper.

Former energy gradients of the meltwater channels need to be established as a primary control on former bed shear stress,flow depth and mean velocity. However, channel gradients measured in the field may only be taken to be equivalent to former energy gradients if the assumption of steady, uniform flow is accepted, *i.e.* assuming that no flow acceleration or deceleration occurred either at-a-station or downstream, and that no rapid instantaneous velocity, largescale eddy or boundary changes were present. Since former steeply graded, coarse-bed meltwater streams are likely to have been subject to rapid velocity fluctuations, to largescale turbulence, and large local energy losses in areas of flow funnelling, expansion and contraction (*e.g.* JARRETT, 1948b), the assumption of steady uniform flow is questionable and acts as a major source of error.

In addition, measurement errors arise in the selection of the length and orientation of the slope measurement in relation to local flow directions, main thalweg directions and valley axis orientation. Where high flow conditions prevailed, as in the channel systems examined in this study, it is likely that former water surface slopes differed little from the overall slope of the terrace surface, although local variability may still have occurred (*e.g.* LORD and KEHEW, 1984; COSTA, 1985).

Errors also arise in attempting to estimate amounts of differential isostatic or tectonic uplift, since in many areas the amounts of rates of uplift in different parts of a terrace system are irregular and have not been accurately determined.

In this study, channel gradients were determined by levelling over distances of up to 200 m in the down-valley direction. Where necessary, adjustments were made for differential isostatic and tectonic uplift of the different terrace surfaces (namely, North Esk, Tekapo and West Greenland), based on published estimates of local uplift rates. The standard deviations of mean slope estimates based both on variability in site selection in relation to channel orientation and downstream location,

TABLE I

Main parameters and sources of error in paleohydrologic analysis of braided outwash terrace sequences

Parameter	Sources of error
A. *Field parameters*	
1. Energy gradient	1. Assumption of steady uniform flow 2. Measurement error of local and overall gradients 3. Estimates of differential isostatic uplift
2. Particle size	4. Selection of sample sites 5. Sample size 6. Measurement procedures and choice of clast diameter 7. Operator error 8. Assumption of maximum sediment availability 9. Assumption of fluvial transport for origin of all sediment
3. Channel pattern, topography and planform	10. Measurement of channel pattern; field levelling of terrace topography. 11. Extrapolation of measures from terrace fragments to whole former outwash surface
B. *Computed hydraulic parameters*	
4. Critical bed shear stress	12. Assumption of validity of Shields critical tractive force approach 13. Value of Shields coefficient
5. Critical flow depth	14. Validity of du Boys equation 15. Substitution of computed values of τ_c and S into du Boys equation 16. Substitution of d_c for R in du Boys equation
6. Flow width	17. Extrapolation of estimated water levels across exposed channel system 18. Assumption of channel contemporaneity 19. Extrapolation of estimated flow width to whole former outwash surface
7. Flow resistance and relative roughness	20. Assumes that particle resistance is dominant source of resistance 21. Small-scale roughness present ($d/p > 3$) 22. Assumption of flow relatively free of suspended sediment 23. Validity of resistance models for n and f, their coefficients and measures of representative grain resistance
8. Critical mean flow velocity	24. Assumption of low or moderate channel gradients 25. Assumes no interaction between channel and overbank flow 26. Assumes validity of paleoflow velocity models
9. Froude number	27. Specification of likely or acceptable range of Froude numbers to validate paleoflow depth and velocity estimates
10. Maximum total discharge	28. Assumes validity of flow continuity equation for steady uniform flow
C. *Independent hydrologic parameters*	
11. Nature of the flow event	29. Assumes preserved channel systems and sediments represent former peak flow events 30. Assumes flow occurred as series of penultimate flood events prior to terrace incision 31. Assumes flow events represented by successive terraced channel systems are directly comparable 32. Assumes terrace incision probably performed by higher magnitude flood events 33. Assumes flow events of terraced channel systems represent minimum flow conditions for terrace incision *i.e.* local threshold conditions for largescale channel change 34. Assumes peak flow events dominated by glacier ablation; partially volcanic eruptions, storm events, glacier bursts 35. Possibility of complex response in terrace formation where differential response in proximal and distal zones
12. Age or date of flow event	36. Lack of accurate and precise dates in some areas 37. Assumes validity of relative and absolute dating methods and accuracy of measurement procedures

and on uplift correction, averaged between 0.007 (Tekapo) and 0.011 (Southwest Greenland).

Particle size characteristics of the outwash deposits provide a means of estimating former skin resistance, and hence former bed shear stress, flow depth and mean velocity. Since the objective was to determine the maximum discharge conditions associated with the formation of each terrace deposit, the coarsest cobbles and boulders present were considered to be of maximum paleohydraulic significance. However, major errors occur in trying to determine accurately the coarse particle size characteristics that are representative of the former flow environment. These errors arise from four sources:

(a) *Selection of sample sites* which should be based on a statistically representative number of randomly located sites. However, site location is often largely controlled by the distribution of exposures, while the number of sites examined is additionally dependent on the time and resources available for analysis. In this study, sample sites were located in all available exposures, and at 20-25 m intervals along a series of traverses across all exposed terrace surfaces and along terrace bluffs. The total number of sites was approximately proportional to the total area represented, ranging between about 40 in West Greenland to nearly 1150 sites in Tekapo).

(b) *Sample size*. The largest size fraction on the bed is considered to have primary control on grain resistance, and hence on eddy intensity and particle entrainment conditions (*e.g.* LIMERINOS, 1970; CHURCH and GILBERT, 1975; CHURCH and KELLERHALS, 1978; DAY, 1978; HOWARD, 1980). Although most traditional flow resistance equations are based on measures of specific coarse percentiles of the whole bed particle size distributions (*e.g.* HENDERSON, 1966; HEY, 1979), a number of workers have used measurements of only the few largest clasts (*e.g.* MALDE, 1968; CLAGUE, 1975; CHURCH, 1978; MAIZELS, 1983a, b, c; LORD and KEHEW, 1984).

The selection of the few largest clasts obviates the need to ensure that a sufficiently large bulk sample has been collected to represent the complete sediment size distribution, a requirement that creates severe practical problems when handling coarse cobble and boulder deposits (*e.g.* Church, 1985, pers. comm.). However, the fraction of the total population that these clasts may represent remains unknown, so that the sample cannot be regarded as being statistically representative of a larger population. This in turn means that the areal limits of a sample site in which clasts are to be measured cannot easily be justified. Nevertheless, in this study, in order to increase areal cover of samples and number of sample sites in the time available, the 10 largest clasts at each site were sampled, with the site extending over an area approximately 10 times the diameter of the largest clast.

(c) *Measurement procedures*. In most hydraulic and paleohydraulic studies the intermediate clast diameter (b-axis) is measured as an indicator of bed roughness (*e.g.* KELLERHALS and BRAY, 1971); this procedure was also adopted in this study. Although LIMERINOS (1970) argued that the minimum diameter (c-axis) is a more valid indication of relative roughness conditions as it represents the height of projection of clasts into the flow, he also found that the choice of axis did not affect the value of the friction factor.

Operator error was minimised in this study since all measurements were completed by one operator (but see HEY and THORNE, 1983). Average standard deviations of clast size samples varied between 0.06 and 0.14 m, representing a percentage variation of from 15.3 to 49.2 per cent of the mean intermediate clast diameter.

(d) *Sediment source and availability*. Paleoflow calculations are also based on the assumption that all sizes of sediment were available for sediment transport and in particular, therefore, that the largest clasts present reflect maximum former stream competence rather than the lack of supply of larger particles to the stream. In the proglacial environments examined in this study this assumption is likely to be acceptable at the catchment scale (*e.g.* HOWARD, 1980), but may not be the case on a local channel scale, especially where bed armouring has occurred, or where fracture systems in the source rocks exert a strong control on clast size (*e.g.* NOLAN and DONNELLY-NOLAN, 1984). In addition, sediment particles are assumed to have been transported by competent Newtonian fluid flows rather than by debris flows, and have not reached the site by ice-rafting, bank slumping, or direct valley-slope inputs (*e.g.* see CHURCH, 1978).

Channel pattern, topography and planform. Measures of channel pattern and local channel relief characteristics are necessary firstly to allow more accurate determinations of (relative and absolute) flow widths and hence total discharges, and secondly to indicate, together with planform changes, the nature of channel responses to changing hydrologic controls. Measures of channel sinuosity, degree of braiding, channel extent and continuity were largely determined in this study from aerial photograph analysis, while local channel relief was determined by levelling of terrace surface cross-profiles. A major source of error arises, however, in the extrapolation of these measures from preserved terrace fragments to the whole former extent of the valley floor.

Age or date of flow event also needs to be determined in relation to each successive terrace surface, as a basis for establishing hydrologic changes through time, as well as estimating possible rates of channel change and terrace formation, and frequency of occurrence of flow events of different relative magnitude. The relative age of the different channel systems investigated in this study has been determined from morphological relationships both with other outwash terraces and with glacial deposits; relative degree of drainage development; thicknesses of loess accumulation; and degree of weathering of clasts. Where only relative dates were available, however, only poor time resolution for paleohydraulic changes was possible.

The absolute dates of formation, or of abandonment by meltwater, of successive paleochannel systems have been determined for the younger terrace sequences of Iceland and Norway. These dates have been obtained by using tephrochronologic and/or lichenometric methods of analysis, often allowing dates of outwash abandonment to be estimated to within ± 2 years (see MAIZELS and PETCH, 1985; MAIZELS and DUGMORE, 1985).

THEORETICAL PROCEDURES FOR PALEODISCHARGE PREDICTION

The theoretical approach adopted in this study for determining former flow conditions in sequences of terraced braided channel systems is based on initial estimates of the critical shear stress at which bed particles begin to move. This method uses SHIELDS' functions (1936; see details in MAIZELS, 1983a), and allows subsequent calculation of former flow depth using du Boys equation, and resistance to flow and flow velocity using Manning and Darcy - Weisbach - type equations. A number of inaccuracies can arise from the assumptions and methods of numerical modeling of paleoflow parameters (*e.g.* ETHRIDGE and SCHUMM, 1978; COSTA, 1983; WILLIAMS, 1984), such that each successive stage of analysis is accompanied by a cumulative expansion in error value (*e.g.* GARDNER, 1983). A primary source or error, for example, is that these procedures all assume that steady, uniform flow conditions prevailed in low relative roughness channels.

Critical bed shear stress has been computed using Shields' function, in which the critical bed shear stress for particle entrainment, τ_c, is expressed as a linear function of particle size, such that

$$\tau_c = \varnothing (\gamma_s - \gamma) D \qquad (1)$$

where Ø is Shields coefficient, γ = specific weight of the fluid (taken as 9810 N m^{-3} for clear, sediment-free water); γ_s = specific weight of the sediment (taken as 26000 N m^{-3}), and D = the representative grain size (in m). The value of Shields coefficient, Ø, is widely taken as 0.056 (*e.g.* see CHURCH, 1978; GRIFFITHS, 1981) but recent work has demonstrated that the value of the coefficient may vary significantly from this figure, depending on the degree of sediment sorting, relative roughness, imbrication, packing and armouring, as well as on particle shape. ASHIDA and BAYAZIT (1973) argue, for example, that at high relative roughnesses critical shear stresses may be underpredicted by Shields diagram by up to 250% because of intense energy dissipation that occurs in separation zones downstream of bed particles. CHURCH (1978) suggests that "underloose" (imbricated) sediments are better represented by coefficients as high as 0.10, and "overloose" ("quick" sediments) by much lower values of *ca.* 0.02. Hence the value of Ø cannot be assumed to be constant for a given deposit, and appears to vary between only 0.01 and as much as 0.25 (WILLIAMS, 1984).

In the present study, therefore, an average Ø value of 0.056 has been adopted (*e.g.* see GRIFFITHS, 1981), but with the recognition that shear stresses may well have been overestimated by up to 30 per cent for many channel deposits (if Ø = 0.04 for example). Such overestimates of Ø would in turn be reflected to a similar degree for flow depth estimates, and in overestimates of mean flow velocity of between 6 and 10 per cent.

Critical flow depth, dc, has been estimated by substituting computed values of τ_c and measured gradients, S, into the du Boys equation for boundary shear, such that

$$d_c = \tau_c / \gamma S \qquad (2)$$

where d_c replaces the value of the hydraulic radius, R, for broad, shallow channels. However, O'BRIEN and RINDLAUB (1934) claim that this substitution may result in shear forces being overestimated by up to 20 per cent and therefore that flow depths may also be overestimated by 4.5 and 7.5 per cent. Substitution of equation (1) into equation (2) gives

$$d_c = C_1 D S^{-1} \qquad (3)$$

where $C_1 = \varnothing (\gamma_s - \gamma)/\gamma$

Flow depths predicted in this way often far exceed the measured channel depths (see MAIZELS, 1983a). These differences probably arise because the sediments were entrained and deposited during overbank flood flows. An empirical test of predicted flow depths using the above approach in an active braided channel system in northern Scotland showed that d_c was predicted with an accuracy of ± 10 per cent. Mean flow depths in the paleochannel systems examined in the present study exhibited large standard deviations, ranging between 12.8 and 74.5 per cent of the overall mean for a given channel system.

Flow width has been estimated by extrapolating estimated water levels, computed from equation 3, across the levelled topographic surface of each channel system. A number of problems arise in the application and interpretation of estimates of flow width. Predicted water levels often varied in successive channels across the terrace surface, but the proportion of channels that were occupied contemporaneously is unknown. In addition, the accuracy of any extrapolation of the proportionate flow width determined for the preserved terrace fragments to the whole extent of the former outwash plain remains uncertain. Hence, at Tekapo, for example, former outwash extents were calculated to have ranged between 1.75 and 12.8 km, whilst flow widths were estimated to have ranged between 0.25 and 1.86 km, respectively, depending on estimates of flow depths, numbers of channels, and surface topography.

Relative roughness and resistance to flow. The accurate determination of mean paleovelocities is largely dependent on obtaining a valid and reliable measure of flow resistance. In low sinuosity, coarse-grained, broad shallow channels such as those characterising the terraced braided channel networks, the dominant source of flow resistance is from bed particle roughness. Where bed roughness is small-scale, *i.e.* where the d/D ratio is above a value of *ca.*3 and where flow is relatively free of suspended sediment (see VANONI and NOMICOS, 1960), skin resistance has been determined from well-established friction coefficients, Manning's n and the Darcy-Weisbach f.

The value of Manning's n is most accurately determined by computation rather than by empirical methods (MAIZELS, 1983a), and in this study 3 different methods of calculating

Manning's have been employed. STRICKLER (1923) derived a simple functional relationship between n and the representative particle diameter such that

$$n = 0.039\ D^{0.16} \quad (4)$$

However, LIMERINOS (1970) demonstrated that n was best determined from a logarithmic function relating the resistance coefficient to a measure of relative roughness:

$$n = \frac{0.113 d^{0.16}}{1.16 + 2.0 \log (d/D_{84})} \quad (5)$$

Where channels are steeply graded with rough beds, an empirical approach for estimating n suggested by JARRETT (1984a) has been used. This equation was based on multiple regression, such that

$$n = 0.32\ S^{0.38}\ d^{-0.16} \quad (6)$$

which was based on slope values of up to 0.034, clast dimensions of up to 0.79 m, and S.E. of 28 per cent.

The Darcy-Weisbach friction factor, f, has, however, a sounder theoretical basis as it is directly related to the von Karman-Prandtl velocity 'law', and is dimensionally correct. In this study, three different methods of computing f have also been selected to provide, together with estimates of n, an envelope of likely resistance coefficients present in the terraced paleochannel systems. Values of f have been determined using the Darcy - Weisbach approach, HEY (1979), in which

$$\frac{1}{\sqrt{f}} = C_2 \log (aR/K_s) \quad (7)$$

where $C_2 = 2.3/(k\sqrt{8})$, where k is the Von Karman coefficient (often taken as 0.04 for clear, sediment-free water); 'a' is a function of cross-sectional channel shape and may be determined either graphically (HEY, 1979) or computed from

$$a = 11.1\ (R/d_{max})^{-0.314} \quad (8)$$

(THORNE and ZEVENBERGEN, 1985); K_s is bed roughness size, taken by HEY (1979) as represented by 3.5 D_{84} (see also BRAY, 1982). However, where the R and/or d_{max} are unknown, f has been estimated from relative roughness alone, according to HENDERSON's (1966) model:

$$f = 0.113\ (d/D)^{0.33} \quad (9)$$

These resistance equations, however, are only valid for small-scale bed roughness channels, with moderate energy gradients. A number of the paleochannels from the braided terrace surfaces exhibit intermediate-scale roughness and relatively steep gradients, often exceeding values of 0.02 m m^{-1}. For such channels, there are not wholly satisfactory resistance equations available that can take account of increased spill resistance, macro turbulence, and 'blocking' effects of large bed roughness elements. Nevertheless, several resistance models have been proposed in the literature (*e.g.* THOMPSON and CAMPBELL, 1979; JARRETT, 1984a; BATHURST, 1985; THORNE and ZEVENBERGEN, 1985), of which THOMPSON and CAMPBELL's (1979) appears to be the most appropriate and has been adopted here, such that

$$\frac{1}{\sqrt{f}} = (1 - 0.1\ K_s/R) \cdot 2 \log (12\ R/K_s) \quad (10)$$

where K_s was found to be best represented by 4.5 D.

Critical mean flow velocity was computed from either the Manning equation:

$$V_c = \frac{d^{0.67}\ S^{0.5}}{n} \quad (11)$$

substituting in values of n determined from equations 4,5 and 6; or the Colebrook-White equation:

$$V_c = \left\{\frac{(8gdS)}{f}\right\}^{0.5} \quad (12)$$

substituting in values of f determined from equations 7, 9 and 10, and where g = acceleration due to gravity (taken as 9.81 m s^{-2}). For steep, rough bed channels paleovelocity was also estimated from an equation derived by COSTA (1983), based on the mean of four theoretical and empirical functions, such that

$$V = 0.18\ (D.\ 10^3)^{0.487} \quad (13)$$

This equation was used by Costa for predicting paleovelocities of flash-floods events with gradients of up to 0.18 m m^{-1} and mean clast sizes of to 1.04 m. Costa found that conventional predictive methods underestimated discharge in these channels by an average of 28 per cent.

The procedures outlined above therefore provided 7 different estimates of critical paleoflow velocity, with scope for extending the estimates according to the choice of Shields coefficient, and to the relative effects of high suspended sediment loads on γ and k, of intermediate rather than short particle axis measurements on relative roughness, and of different percentile or clast number measures of particle size on the resistance coefficients. These additional effects and models available suggest that the paleovelocities determined using the current procedures are unlikely to be highly accurate; they are, however, likely to represent the correct order of magnitude, and to reflect actual overall patterns of longterm change in relative flow magnitude.

Froude numbers provide an important test of the likely validity of the paleoflow depth and velocity estimates. Froude numbers appear rarely to exceed values of about 2.5, and often lie below the critical value of 1.0 in steep, coarse-grained braided channels (*e.g.* JARRETT, 1984a; BATHURST, 1985; COSTA, 1985; THORNE and ZEVENBERGEN, 1985) and hence have provided an upper limit on the range of values likely to be valid for the paleochannel systems examined in this study.

Maximum total discharge has been estimated from the flow continuity equation,

$$Q = w.d.V_c \quad (14)$$

The total discharges estimated in this study have been based on the product of (a) estimates of flow width determined by extrapolation of estimated flow depths across preserved channel cross-sections and extended across the complete former outwash surface, and from measurements of the num-

TABLE II

Input parameters and estimates of selected paleoflow parameters for deglaciation terrace sequences (mean values for each parameter given for Shields coefficient of 0.056)

Terrace deposit	Approx. age	Paleoflow parameters: Gradient (m m^{-1})	Clast size (m)	Flow width (m)	Critical tractive force (N m^{-2})	Flow depth (m)	Relative roughness	(Mean values for Ø = 0.056) n-value	f-value	Velocity (m s^{-1})	Froude No.	Discharge: Q1	Q2	Q3	Q4	Q5	Q6	Q7
North Esk, N.E. Scotland																		
T3	*ca.* 15 500 yrs BP	0.0050	0.042	1688.0	38.25	0.78	18.48	0.032	0.172	1.79	0.54	3 430	2 687	1 773	3 423	1 333	1 466	(7 945)
Tekapo, S. Island, New Zealand																		
MJ1		0.0112	0.207	979.0	187.86	1.71	8.25	0.042	0.146	3.57	0.92	8 450	6 136	4 755	8 190	4 296	4 047	13 326
MJ2		0.0099	0.218	926.0	198.01	2.03	9.29	0.040	0.149	3.77	0.87	9 929	7 312	6 060	9 682	4 853	4 659	16 638
MJ3	18 000-16 000	0.0099	0.256	844.0	232.46	2.38	9.29	0.041	0.149	4.12	0.88	11 512	8 477	7 404	11 227	5 626	5 390	(19 290)
MJ4	yrs BP	0.0082	0.214	561.0	193.57	2.39	11.20	0.038	0.154	3.88	0.79	7 235	5 435	4 849	7 116	3 322	3 291	(13 392)
MJ5		0.0154	0.259	661.0	235.09	1.55	5.98	0.046	0.139	3.77	1.12	5 486	3 812	2 793	5 212	3 105	2 765	(6 959)
Tek 1		0.0223	0.323	856.0	292.39	1.34	4.15	0.051	0.186*	4.12*	1.40*	6 439	4 199	2 891	5 935	4 116	3 439	6 084
Tek 2		0.0134	0.243	102.0	220.22	1.68	6.90	0.044	0.141	3.75	1.03	906	643	487	872	489	446	(1 268)
Tek 3	15 000-13 000	0.0073	0.242	403.0	219.59	3.05	12.57	0.038	0.157	4.27	0.76	7 186	5 458	5 343	7 104	3 174	3 201	(14 077)
Tek 4	yrs BP	0.0074	0.278	282.0	252.05	3.47	12.49	0.038	0.157	4.61	0.77	6 152	4 670	4 768	6 086	2 724	2 739	(12 013)
Tek 5		0.0076	0.249	250.0	226.03	3.03	12.16	0.038	0.156	4.31	0.77	4 477	3 390	3 300	4 424	2 000	2 005	(8 630)
Søndre Strømfjord, W. Greenland																		
T1		0.0117	0.177	205.0	160.48	1.40	7.90	0.035	0.144	3.25	0.94	1 327	959	693	1 285	685	642	(2 024)
T2*		0.0259	0.303	355.3	274.71	1.08	3.57	0.054	0.203	3.78	1.48	2 037	1 287	825	1 850	1 370	1 117	1 664
T3*	7500-6500 yrs BP	0.0204	0.306	380.0	277.16	1.38	4.53	0.050	0.178	4.14	1.35	2 919	1 935	1 349	2 713	1 813	1 537	2 977
T4		0.0032	0.240	23.2	217.23	6.92	28.88	0.032	0.191	5.26	0.55	1 074	863	1 247	1 109	360	417	(2 920)
T5		0.0204	0.313	28.3	283.51	1.42	4.53	0.050	0.177	3.42	1.36	225	149	105	211	140	118	229
T6	6500-5500 yrs BP	0.0140	0.148	36.2	134.36	0.98	6.60	0.044	0.140	2.85	1.03	147	103	65	140	80	73	(198)
T7	700-300 yrs BP	0.0131	0.092	552.7	83.68	0.65	7.06	0.043	0.142	2.24	0.97	1 181	840	469	1 134	633	587	(1 677)
Narssarssuaq, S.W. Greenland																		
T1*		0.0235	0.410	8.8	371.90	1.61	3.93	0.052	0.190	4.62	1.46	89	57	42	84	58	48	80
T2*		0.0641	0.515	612.9	467.10	0.74	1.44	0.075	0.253	3.67	2.33	2 708	1 299	800	2 156	2 463	1 715	570
T3*	2500-2250 yrs BP	0.0437	0.541	285.8	490.76	1.14	2.11	0.065	0.329	4.26	1.90	2 126	1 173	785	1 809	1 702	1 263	907
T4*		0.0262	0.469	64.9	425.40	1.66	3.53	0.054	0.204	4.74	1.51	708	446	329	646	478	387	572
T5		0.0175	0.460	41.7	416.60	2.43	5.28	0.048	0.136	5.01	1.25	706	481	405	672	417	361	(816)
T6*		0.0290	0.430	59.2	390.22	1.37	3.19	0.056	0.219	4.37	1.57	504	311	215	454	352	280	336
T7	2250-2200 yrs BP	0.0233	0.453	53.8	410.71	1.80	3.97	0.052	0.190	4.86	1.45	638	413	312	590	414	342	(578)
T8*	Post-1880 AD	0.0215	0.245	155.0	221.85	1.05	4.30	0.051	0.182	3.61	1.37	802	527	335	743	507	427	782
T9*		0.0247	0.153	291.0	138.35	0.57	3.74	0.053	0.197	2.68	1.42	631	403	210	576	417	346	540
Solheimajökull, S. Iceland																		
T1	Pre 630 AD	0.0151	0.222	3280.0	201.27	1.36	6.12	0.045	0.139	3.90	1.20	22 152	15 443	10 851	21 070	12 443	11 142	(28 559)
T2*	~ 1000	0.0263	0.103	1440.0	93.16	0.42	4.07	0.052	0.189	2.22	1.41	1 881	1 221	578	1 730	1 213	1 170	1 742
T3*	~ 1000	0.0263	0.513	1440.0	465.26	1.83	3.65	0.054	0.208	4.83	1.51	17 590	11 102	8 594	15 984	11 847	9 916	14 452
T4	~ 1250	0.0132	0.207	2005.0	187.67	1.45	7.00	0.043	0.142	3.45	1.01	14 263	10 138	7 343	13 689	7 660	7 021	(20 150)
T5	~ 1300	0.0128	0.276	1370.0	250.23	1.99	7.22	0.043	0.143	4.05	1.01	15 554	11 102	8 944	14 958	8 268	7 589	(22 416)
T6*	~ 1350	0.0260	0.310	1310.0	281.06	1.10	3.55	0.054	0.204	3.82	1.48	7 743	4 890	3 154	7 026	5 213	4 246	6 301
T7*	~ 1500	0.0400	0.469	1820.0	425.21	1.08	2.31	0.063	0.296	4.08	1.81	12 108	6 858	4 476	10 424	9 412	7 097	5 883
T8	~ 1840	0.0134	0.284	1230.0	257.49	1.96	6.90	0.044	0.142	4.64	1.17	13 817	9 800	7 825	13 251	7 458	6 791	(19 329)
T9	~ 1953	0.0125	0.183	145.0	165.92	1.35	7.39	0.043	0.143	3.26	0.97	914	654	465	881	482	446	(1 337)
T10	~ 1937	0.0121	0.144	655.0	130.56	1.10	7.64	0.042	0.144	2.89	0.94	2 992	2 152	1 434	2 888	1 561	1 459	(4 469)
T11	~ 1935	0.0119	0.208	595.0	188.58	1.62	7.77	0.042	0.144	3.53	0.96	4 812	3 468	2 624	4 649	2 497	2 328	(7 260)
T12	~ 1969	0.0117	0.112	510.0	101.54	0.88	7.90	0.042	0.144	2.54	0.91	1 162	1 200	747	1 607	858	808	(2 534)
T13	~ 1883	0.0127	0.260	410.0	235.73	1.89	7.28	0.043	0.143	3.93	1.00	4 295	3 069	2 433	4 135	2 277	2 095	(6 221)
T14	~ 1884	0.0104	0.244	70.0	221.22	2.17	8.89	0.041	0.147	3.97	0.90	842	617	520	822	418	397	(1 376)
T15	~ 1935	0.0119	0.305	135.0	276.53	2.37	7.77	0.043	0.144	4.35	0.98	1 939	1 397	1 198	1 878	1 006	933	(2 925)
T16*	~ 1937	0.0201	0.310	125.0	281.06	1.43	4.60	0.050	0.177	4.18	1.35	998	664	469	931	617	525	1 031
T17	~ 1949	0.0116	0.366	185.0	331.38	2.91	7.97	0.043	0.144	4.80	0.98	3 561	2 575	2 375	3 454	1 832	1 703	(5 458)
T18	~ 1969	0.0116	0.270	100.0	244.79	2.15	7.97	0.042	0.144	4.10	0.96	1 232	891	742	1 196	634	592	(1 888)
Austerdalen, S. Norway																		
T1	~ 1756 AD	0.0170	0.140	165.0	126.57	0.77	5.49	0.047	0.137	2.66	1.15	491	337	195	465	286	253	(584)
T2*	~ 1753	0.0254	0.149	163.0	146.04	0.84	5.63	0.052	0.216	3.06	1.46	558	371	220	518	342	282	643
T3*	~ 1769	0.0250	0.164	79.0	148.69	0.64	3.89	0.053	0.201	2.81	1.43	200	128	70	183	131	109	176
T4	~ 1821	0.0131	0.086	141.0	77.70	0.60	7.06	0.043	0.142	2.16	0.97	270	192	104	257	143	133	380
T5	~ 1794	0.0187	0.103	185.0	93.47	0.52	5.01	0.048	0.136	2.23	1.20	315	212	108	295	189	166	348
T6*	~ 1850	0.0379	0.168	130.0	152.22	0.41	2.44	0.062	0.279	2.43	1.72	198	114	54	172	151	116	103
T7*	~ 1842	0.0331	0.137	93.0	124.21	0.38	2.80	0.059	0.245	2.29	1.61	121	72	33	107	88	70	76
T8*	~ 1848	0.0239	0.149	93.0	156.38	0.64	4.33	0.052	0.189	2.87	1.46	237	155	82	219	150	122	232
T9	~ 1833	0.0190	0.130	81.0	118.04	0.67	5.15	0.048	0.137	2.54	1.28	200	135	75	188	120	105	224
T10	~ 1868	0.0092	0.101	94.0	91.84	1.34	13.23	0.038	0.158	2.63	0.78	472	354	269	466	215	215	908
T11*	~ 1870	0.0251	0.132	161.3	119.68	0.54	4.07	0.053	0.207	2.54	1.43	310	199	103	284	203	170	283
Nigardsdalen, S. Norway																		
T1*	~ 1802 AD	0.0231	0.117	165.0	106.26	0.47	4.04	0.052	0.190	2.40	1.38	261	169	83	240	169	142	240
T2*	~ 1803	0.0277	0.183	46.0	165.46	0.61	3.34	0.055	0.212	2.83	1.50	114	71	38	104	79	64	87
T3*	~ 1830	0.0217	0.129	33.5	117.23	0.59	4.57	0.050	0.186	2.63	1.35	71	47	25	66	45	38	72
T4*	~ 1791	0.0284	0.179	84.0	162.47	0.58	3.26	0.056	0.216	2.78	1.52	196	97	64	177	136	110	146
T5*	~ 1807	0.0301	0.212	51.0	192.03	0.76	3.61	0.056	0.240	3.10	1.56	170	106	61	154	117	95	137
T6*	~ 1808	0.0287	0.150	57.0	136.36	0.49	3.23	0.056	0.218	2.53	1.52	103	63	31	93	71	58	76
T7*	~ 1821	0.0259	0.208	38.0	188.22	0.78	3.74	0.054	0.206	3.14	1.47	131	83	48	120	87	72	111
T8*	~ 1826	0.0272	0.195	32.0	176.70	0.67	3.44	0.055	0.210	2.96	1.50	91	57	31	83	62	50	71
T9*	~ 1836	0.0277	0.175	152.0	158.57	0.59	3.36	0.055	0.213	2.77	1.50	358	223	117	323	246	200	275
T10*	~ 1843	0.0240	0.178	59.0	161.20	0.68	3.85	0.052	0.193	2.94	1.42	165	106	59	152	108	90	145
T11*	~ 1844	0.0271	0.169	130.0	153.40	0.58	3.45	0.055	0.210	2.75	1.48	297	186	98	269	202	165	234
T12*	~ 1886	0.0195	0.130	103.0	118.04	0.62	4.74	0.049	0.136	2.49	1.24	233	156	83	217	142	123	(247)
T13	~ 1890	0.0155	0.129	68.0	117.23	0.77	5.96	0.045	0.139	2.60	1.08	198	137	80	188	112	101	(251)

Key to discharge estimates. Velocity based on resistance equations: Q1 Eqn. (4) (STRICKLER, 1923); Q2 Eqn. (5) (LIMERINOS, 1970);

bers of channels and their widths across preserved terrace surfaces; (b) estimates of flow depth based on equation (3), using an average Shields coefficient of 0.056; and (c) estimates of flow velocity based on equations (11) to (13).

This procedure compounds the many sources of error that originate in the determination of flow widths, depths and velocities. Detailed error analysis has not yet been completed, but the estimates of mean paleodischarge exhibit high variability reflecting the range of possible width, depth and velocity estimates for each paleochannel system. Estimates of the variability in mean paleodischarge values have been calculated using measures of ± 1 standard deviation from the mean width, depth and velocity estimates for the individual channel systems. Mean discharge values were found to vary on average by ± 12% (range 7 to 16%), ± 22% (range 11 to 41%), and ± 33% (range 13 to 92%) according to the standard deviation of the velocity, width and depth estimates, respectively. Hence, single mean discharge values were found to vary on average by ± 53% (range 36 to 93%) according to the combined standard deviations of width, depth and velocity estimates, while the range of discharge values predicted by the seven different models (Q1 to Q7 in Table II) varied on average by an additional 40%.

EVIDENCE OF PALEODISCHARGE CHANGE DURING DEGLACIATION

The modeling procedures have provided minimum and maximum limits for all the major paleoflow parameters, including flow depths (based on different values of Shields coefficient), resistance measures, velocities, Froude numbers, and discharges. *Although much of the data is summarised in Table* II, the scope of this paper only allows for brief analysis of the results of the paleodischarge estimates for each of the terrace sequences.

North Esk, northeast Scotland. The North Esk outwash plain, lying southeast of the Grampian Mountains, exhibits a sequence of 4 main terrace surfaces dating from the Late Devensian ice sheet retreat *ca.* 15,500-14,500 years BP. The different terrace surfaces are marked by extensive braided paleochannel systems (MAIZELS, 1983c). The area is extensively cultivated and wooded and few natural exposures are available. Hence paleodischarge estimates have been made for the one most distinctive surface only (T3, MAIZELS, 1983c).

The paleodischarge estimates, depicted on Figure 2 as the minimum and maximum limits of the different predictive models, indicate an overall decrease in peak discharges of 10 to 30 times from the Late Devensian surface compared with the present day flood record (Tay River Purification Board, pers. comm.). However, this evidence is not very helpful in examining progressive changes through time, as represented by a complete terrace sequence.

Tekapo valley, South island, New Zealand. The Tekapo valley is marked by a sequence of 10 extensive outwash terraces dating from the Wolds glaciation (*ca.* 105,000 years BP), the Balmoral glaciation (*ca.* 51,000 years BP), and more recent terrace sequences of the Mt. John (*ca.* 16,000 years BP) and Tekapo (*ca.* 14,000 years BP) glacial phases (SPEIGHT, 1961; GAIR, 1967; MANSERGH, 1973), each glacial phase being of successively lesser extent. The older two deposits are mantled by thick loess sediments (*ca.* 2.5 m thick), partially masking the original outwash channel patterns, and reducing the availability of sediment exposures. The younger two Lateglacial terrace sequences, by contrast, are marked by complex and extensive braided paleochannel networks, in which the sediments are widely exposed both on the terrace surfaces and along the terrace bluffs.

The estimated paleodischarges associated with the two Late Glacial terrace sequences of Mt. John and Tekapo appear to have reached a peak during formation of the early and middle terraces, declining towards the next interstadial or post-glacial phase (Fig. 3). The higher terraces of the smaller, later Tekapo glacial episode appear in turn to have been associated with lower peak discharges than those of the equivalent Mt. John terraces. The estimated paleodischarges exceed the maximum recorded flood flow of 641 $m^3 s^{-1}$ (Ministry of Works & Development, Wellington, N.Z., *pers. comm.*) (estimated as representing a 2500-year flood), by up to 26 times. However, the poor temporal resolution, the large overlap in the range of estimated paleodischarges for successive

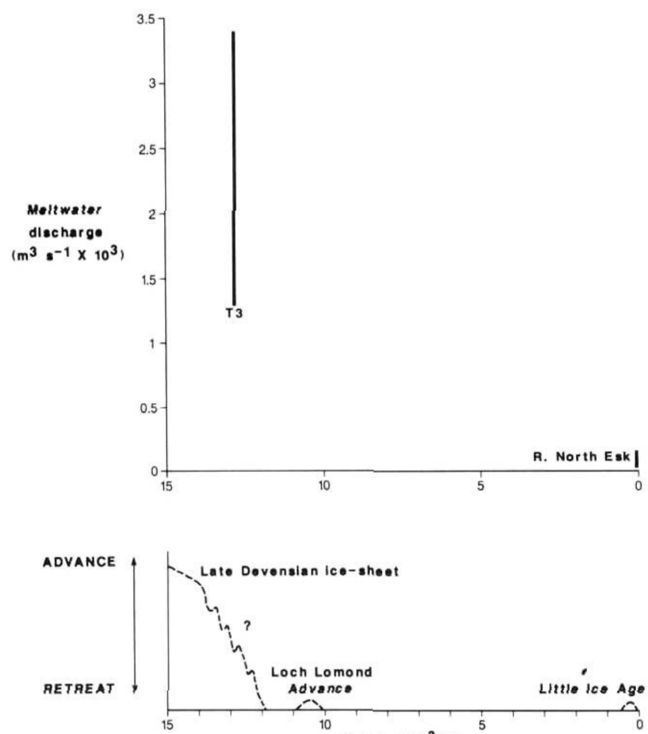

FIGURE 2. Paleohydrologic change during deglaciation from T3 outwash surface, North Esk, northeast Scotland (vertical discharge bars represent range of minimum and maximum values estimated from Q1 to Q7).

Changements paléohydrologiques survenus au cours de la déglaciation, observés à partir de l'épandage fluvio-glaciaire T3, North Esk, nord-est de l'Écosse; la ligne verticale (écoulement de l'eau de fonte) représente l'écart entre les valeurs maximales et minimales estimées de Q1 à Q7.

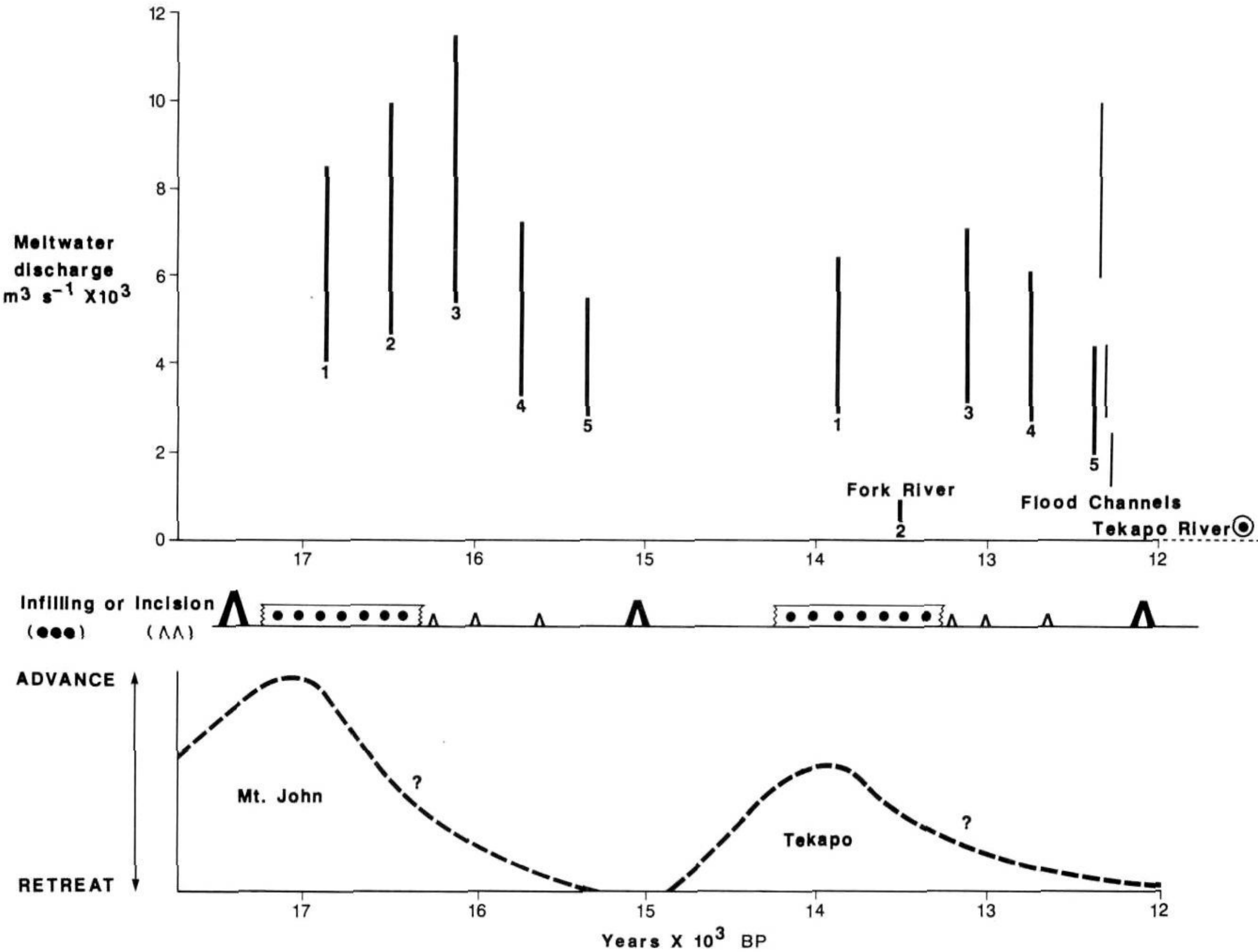

FIGURE 3. Paleohydrologic change during deglaciation. Mt.John and Tepako outwash terraces, South Island, New Zealand.

Changements paléohydrologiques survenus au cours de la déglaciation, terrasses fluvio-glaciaires du Mt. John et de Tekapo, South Island, Nouvelle-Zélande.

deposits, and the inadequacy of data on the glacial advances and retreats, preclude the definition of any convincing pattern of hydrologic change during deglaciation.

Sondre Strømfjord, West Greenland. A series of eight terrace forms bound the valley of the Watson River about 12 km from the present western margin of the Greenland ice sheet (MAIZELS, 1983b). Although several of these terrace features are narrow and fragmented, they do exhibit distinctive evidence of paleochannel courses. This area is believed to have become deglaciated between *ca.* 7000 and 6500 years BP, following the Keglen stage 're-advance' (WEIDICK, 1968, 1976; TEN BRINK, 1975). Maximum deglaciation was attained about 6000 yr BP, followed by a further small re-advance, while isostatic uplift of up to 40 m is believed to have ceased by about 4000 yr BP. During the latter part of the Holocene, only small-scale ice marginal oscillations have occurred, culminating in the Little Ice Age, or Ørkendalen, re-advance, between 700 and 300 yr BP.

Paleodischarge estimates for the seven most distinctive terrace deposits suggest that discharges were highest during the period of most rapid deglaciation (channel systems 2 and 3, Fig. 4). At the earliest stage of deglaciation (channel system

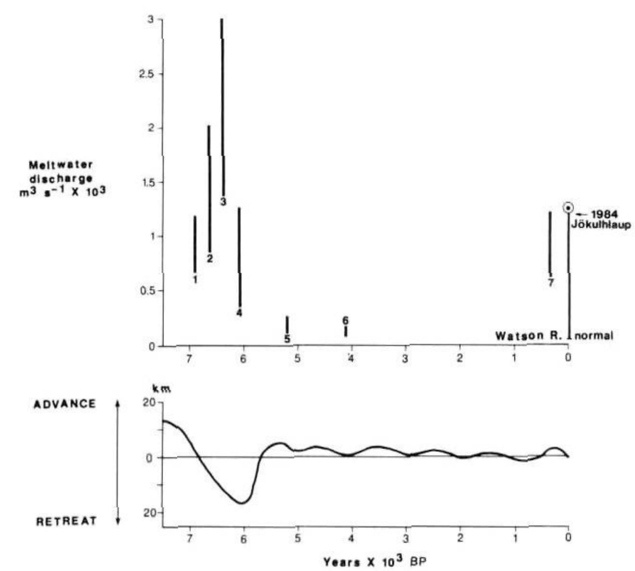

FIGURE 4. Paleohydrologic change during deglaciation, Søndre Strømfjord, west Greenland.

Changements paléohydrologiques survenus au cours de la déglaciation, Søndre Strømfjord, ouest du Groenland.

1), and throughout the remaining period of the Holocene, by contrast, peak discharges have remained comparatively low. Nevertheless, even the highest peak paleodischarges are less than 3 times the maximum recently recorded jökulhlaup peak (1984 flood, Fig. 4; SUGDEN *et al.*, 1985). Hence, although distinctive contrasts appear to have occurred in association with largescale ice melt in the early Holocene, the continued proximity of the ice sheet during the Holocene has acted to maintain peak meltwater discharges at a similar order of magnitude throughout this period. The poor dating framework and time resolution available for these terrace systems unfortunately remain too crude to allow more precise interpretation.

Narssarssuaq, S. W. Greenland. The Narssarssuaq proglacial valley extends for some 8 km from the snout of the Kiagtut sermia glacier to the estuary of the meltwater stream where it enters Eriksfjord in south-west Greenland. Of six main terrace systems in the valley, the oldest (T5; MAIZELS, 1983b) terrace includes deposits that appear to date from the mid-Sub Atlantic ice recession *ca.* 2500-2200 yr BP (WEIDICK, 1963). The terrace is about 600 m wide, and exhibits a range of channel types (T1 to T7, Table II) at different elevations across its surface. The highest, oldest channels are proximal braided forms, steeply graded and extensively pitted with kettleholes (T1 to T3); the lower, more recent channels become increasingly broad, deep and sinuous (T4 to T7).

Estimates of paleodischarge reflect this clear pattern of channel change, apparently representing the incision and migration of increasingly sinuous streams during deglaciation. Paleodischarges exhibit maximum peak values in association with the earlier deposits during initial deglaciation, followed by an abrupt decrease to consistently low discharges over the remaining period of deglaciation and glacial fluctuation (Fig. 5).

Solheimajökull, southern Iceland. The Solheimajökull glacier drains southward for *ca.* 8 km from Myrdalsjökull ice cap, which is underlain by the volcano Katla. Hence the longterm *meltwater drainage changes in the Solheimajökull valley may* be related not only to changes in seasonal ablation and meltwater yields, but also to periodic catastrophic floods (or jökulhlaups) that are produced during volcanic eruptions. The occurrence of volcanic eruptions has also resulted in the widespread accumulation of tephra layers; these are readily identified in sediment and soil sequences, and have provided a relatively precise dating framework for the terrace deposits in the Solheimajökull area for the past 1500 years. Older deposits cannot be precisely dated, but have been distinguished on stratigraphic and weathering criteria. In addition, the time resolution available for deposits dating from the Little Ice Age maximum, between *ca.* 1850 and 1890 AD, has been significantly improved through lichenometric dating methods (MAIZELS and DUGMORE, 1985). Individual deposits have been dated to within ± 2 years of their being abandoned by *meltwaters*.

Twenty different meltwater terrace deposits have been identified in the Solheimajökull valley, including several distinctive jökulhlaup channel deposits related to eruptions of Katla *ca.* 1000 AD. The estimated paleodischarge changes since this time exhibit a longterm overall decline associated with ice retreat and subsequent small scale fluctuations (Fig. 6), punctuated by 2 major jökulhlaup peaks prior to *ca.* 1000 yrs BP. Little Ice Age discharges have remained relatively low and similar in magnitude to those estimated for the present day meltwater river; Solheimajökull, however, has retreated only *ca.* 1.3 km from its Little Ice Age limit.

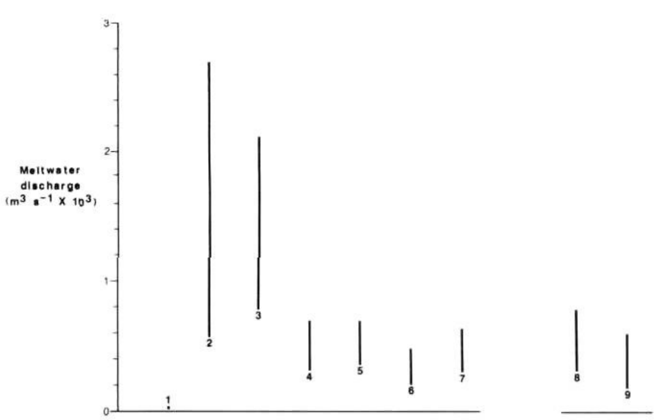

FIGURE 5. Paleohydrologic change during deglaciation, Narssarssuaq, southwest Greenland.

Changements paléohydrologiques survenus au cours de la déglaciation, Narssarssuaq, ouest du Groenland.

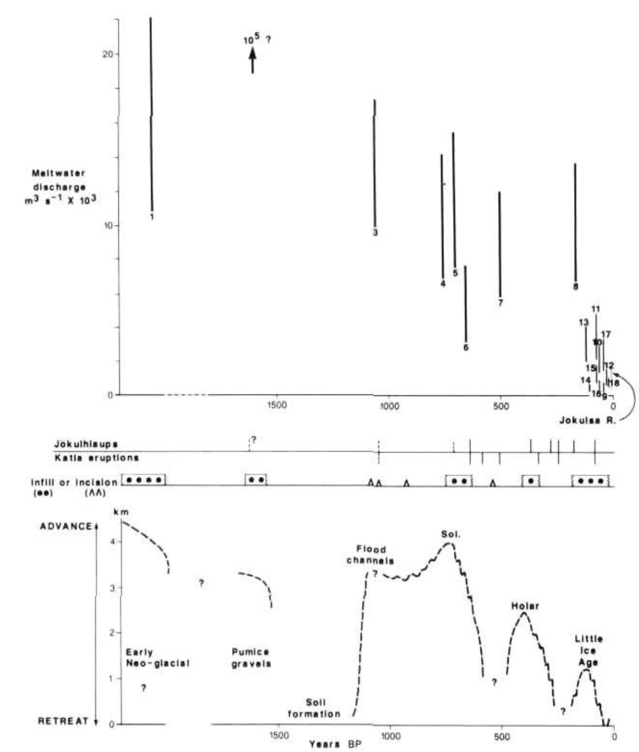

FIGURE 6. Paleohydrologic change during deglaciation, glacier fluctuations and volcanic eruptions, Solheimajökull, southern Iceland.

Changements paléohydrologiques survenus au cours de la déglaciation, fluctuations glaciaires et éruptions volcaniques, Solheimajökull, sud de l'Islande.

Nigardsdalen and Austerdalen, southern Norway. The upper reaches of these two valleys are occupied by outlet glaciers from the Jostedalen ice cap, having receded from their Little Ice Age maxima of about 1748-1750 AD. They left in their wake a series of terminal moraine ridges, separated by sequences of braided outwash deposits. The outwash deposits have been dated to within ± 2 years by lichenometric methods (MAIZELS and PETCH, 1985), based on a well-established *Rhizocarpon geographicum* growth curve for the area (ANDERSON and SØLLID, 1971). The braided channel systems are distinctive on each surface, and clast deposits are widely exposed.

Estimates of paleodischarge change at Nigardsdalen for the period of glacier retreat between *ca.* 1800 and 1900 AD show no significant trend through time. The pre-1800 deposits have been buried by later outwash, thereby removing evidence for initial deglaciation flows. However, the remaining discharge fluctuations appear to reflect the estimated fluctuations in rates of ice retreat, with peaks occurring around 1830-1860 AD (Fig. 7a).

Estimates of paleodischarges at Austerdalen, by contrast, appear to exhibit an overall marked decline following the initial period of rapid ice retreat between *ca.* 1750 and 1760 (Fig. 7b). As at Nigardsdalen, periodic discharge peaks appear to be related to periods of more rapid net ice retreat, occurring at the beginning and end of the terrace sequence (channel systems 1, 2 and 10, 11, respectively; Fig. 7b). However, the temporal resolution for determining rates of ice retreat is too crude to allow direct correlation with paleodischarge estimates, since any short-term ice-marginal fluctuations have been masked by the longer term average rates of ice retreat.

DISCUSSION AND CONCLUSIONS

The estimates of paleodischarge for the individual channel systems largely represent a plausible range of flows that compare closely with meltwater flows recorded in many active proglacial channel systems (Table II; and see CHURCH and GILBERT, 1975; CLAGUE, 1975). In particular, the predicted flow depths (0.38 to 6.92 m), resistance coefficients ($n = 0.032$ to 0.075; $f = 0.136$ to 0.329), velocities (1.79 to 5.26 m s^{-1}), and Froude numbers (0.54 to 2.33) closely match those recorded in many active braided meltwater channels (*e.g.* BOOTHROYD and ASHLEY, 1975; RICE, 1982). These results suggest that reasonable confidence may be placed in the relative magnitude of the different paleodischarge estimates.

When examining the temporal changes in the peak flows occurring through each terrace sequence, it appears that peak paleodischarges have declined only in areas experiencing longterm or largescale deglaciation. On shorter timescales, any trend that might exist has been disrupted by significant

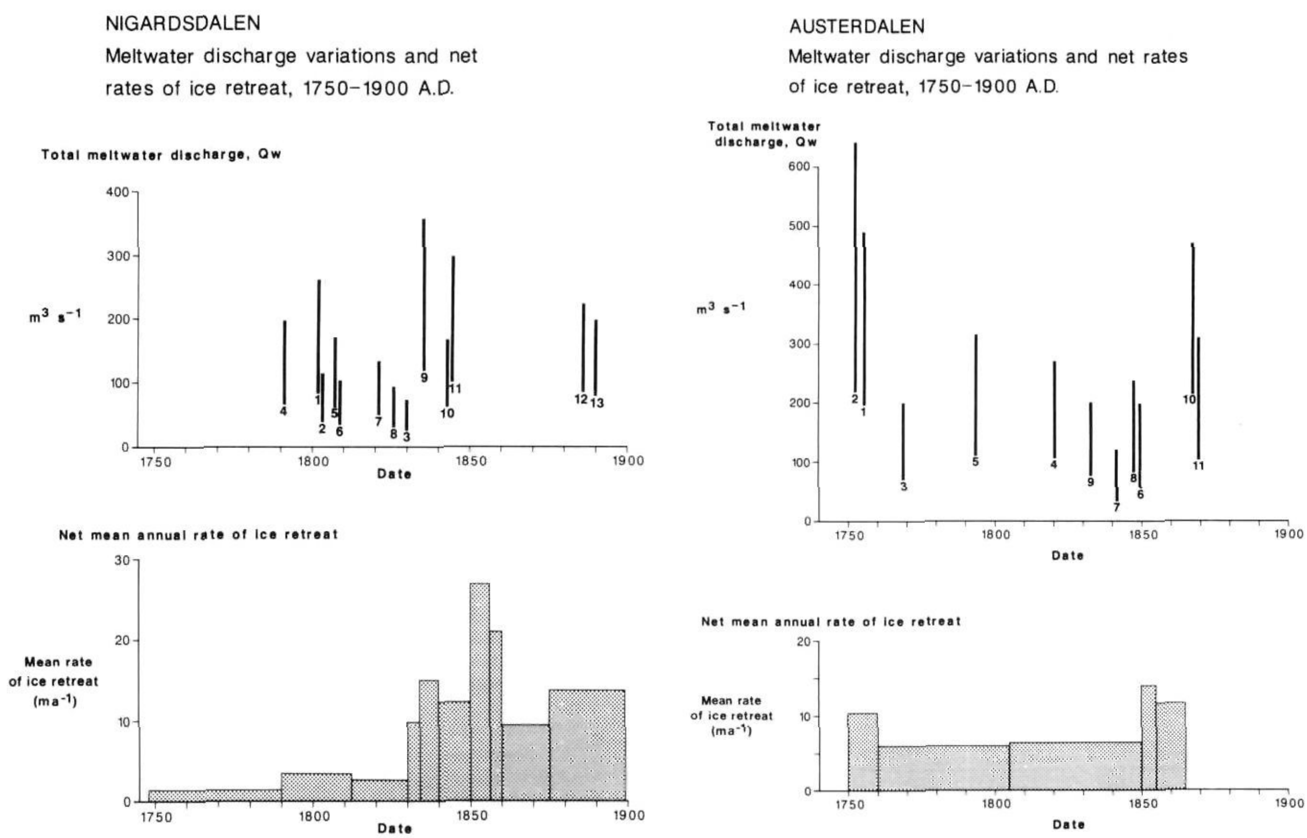

FIGURE 7. Paleohydrologic change during ice retreat, Jostedalsbreen: a) Nigarsdalen outwash deposits; b) Austerdalen outwash deposits.

Changements paléohydrologiques survenus au cours du retrait glaciaire, Jostedalsbreen: a) dépôts fluvio-glaciaires de Nigardsdalen; b) dépôts fluvio-glaciaires d'Austerdalen.

short-term fluctuations and episodic flood events, produced during short periods of rapid ice melt, by storm precipitation events, or by jökulhlaup drainage.

Thus, in the North Esk catchment where ice has since disappeared, maximum discharge changes of up to 30 times have been recorded in association with significant channel changes (MAIZELS, 1983c). In catchments where ice has receded over a longterm time-scale (*e.g.* Tekapo, W. Greenland, S. W. Greenland, Iceland) but still remains in the catchment, paleodischarges appear to have reached a maximum during periods of most rapid ice melt, apparently during the early stages of deglaciation, and with lower, but still highly variable, peak flows occurring during subsequent periods of deglaciation. Maximum discharge changes over the different timescales recorded ranged from 39.8 (Tekapo), 2.4 (Søndre Strømfjord), 4.3 (Narssarssuaq) and 18 (Solheimajökull) times (minimum equivalents are 19, 1.25, 1.24 and 9 times, respectively). The high variability of flows recorded in paleochannel systems associated with periods of shorter term deglaciation (*e.g.* Norway), also appears to mask any overall trend in peak discharges, as a result of both melt and non-melt mechanisms acting to produce peak flow events.

A model of overall discharge increase followed by decline during progressive deglaciation does appear to be acceptable over longterm timescales or over large spatial scales, where significant proportions of ice have disappeared from the catchment. On a shorter timescale, however, local controls on short-term melt rates and on non-melt processes, such as volcanically-induced floods, play a dominant role in channel development, and any longer term trends in peak discharges are obscured.

The development of such a model, incorporating both "longterm" and "shortterm" paleohydraulic changes during deglaciation must remain in a tentative stage until the overall approach, the assumptions and the modeling procedures adopted can be fully validated. In particular, the methodology employed in deriving the field parameters needs to be tested rigorously both in terms of sampling and measurement procedures, and in terms of their application in the models used. The paleohydraulic models themselves need to be thoroughly assessed in terms both of their validity as theoretical representations of critical flow conditions and of their applicability to the paleo-environments being studied. Alternative models of paleo-flow need to be examined and tested in similar ways, especially in analogous present-day environments (*e.g.* see discussion in SCHUMM, 1985).

The procedures and models of discharge change need to be tested and applied to a wider range of meltwater channel environments, including erosional sedimentary channels and alluvial spillways. The approach also needs to be extended to the analysis of complete sedimentary sequences that underlie the terrace surfaces, rather than relying on relatively few of the total flow events represented by a complete valley fill. Finally, the estimates of discharge change based on paleohydraulic modeling need to be tested against flow estimates obtained from independent evidence, such as glaciological models, slack-water deposits or sediment records of meltwater yields (*e.g.* KOCHEL and BAKER 1982; BAKER *et al.*, 1983; WAITT, 1985; Sundborg, pers. comm.) and in areas with improved dating resolution (*e.g.* KNOX, 1985).

Nevertheless, despite the many sources of error and uncertainty, with further work along the lines discussed above, this paleohydraulic approach to the analysis of proglacial terraced channel systems could prove to be an important tool for determining hydrologic changes during deglaciation in particular, and for more extensive paleo-environmental reconstructions in general.

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