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présentée par :

Timothée Jautzy

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pour obtenir le grade de : **Docteur de l'université de Strasbourg**

Discipline/S spécialité : Géographie – Géomorphologie

Dynamique morpho-sédimentaire des rivières vosgiennes. Emboîtement d'échelles spatio-temporelles pour un diagnostic géomorphologique fonctionnel.

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A Nico...

Avant-Propos

Ce travail de thèse s'inscrit dans le cadre du projet de recherche intitulé SEDIVOSGES, qui vise à caractériser la dynamique morpho-sédimentaire des rivières vosgiennes à des échelles spatio-temporelles emboîtées, dans l'objectif d'établir un diagnostic géomorphologique fonctionnel de la morphodynamique des hydrosystèmes fluviaux vosgiens. Ce travail est à la fois d'ordre fondamental et appliqué. Les nouvelles connaissances fondamentales acquises doivent venir en soutien de la gestion appliquée des rivières vosgiennes.

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Cette thèse est présentée sous la forme de trois articles, en langue anglaise, dont le premier est publié et les deux autres sont en cours de soumission. En complément de ces articles, un chapitre d'introduction, rédigé en français, vient problématiser les enjeux et les hypothèses de ce travail. Pour finir, un chapitre de synthèse, également rédigé en français, offre une vue d'ensemble des résultats et des perspectives futures.

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Résumé

La dynamique morpho-sédimentaire dans le massif des Vosges reste mal connue. Cette lacune touche autant les flux sédimentaires aux échelles temporelles actuelle et subactuelle que les trajectoires temporelles longues des stocks sédimentaires des fonds de vallées. Or, la temporalité longue est fonctionnellement dépendante de la temporalité courte. Ainsi, l'élaboration de stratégies de gestion, voire de restauration, efficaces et durables des systèmes fluviaux, doit impérativement tenir compte de la temporalité longue pour mieux comprendre la dynamique morpho-sédimentaire actuelle. Le massif des Vosges présente trois caractéristiques majeures qui en font un objet particulièrement propice à l'étude et à la comparaison de la dynamique morpho-sédimentaire de ses systèmes fluviaux, dans une approche systémique, à échelles spatio-temporelles emboîtées. D'abord, l'origine géologique de sa formation explique de forts contrastes lithologiques, morphométriques et climatiques. Ensuite, il présente la rare particularité de n'avoir été englacé que partiellement. Enfin, le massif des Vosges est le plus densément peuplé de France, ce qui constitue à la fois un facteur de contrôle potentiel de la dynamique morpho-sédimentaire moderne, dont il convient de déterminer l'ampleur, et un enjeu en termes de gestion des hydrosystèmes. Ce travail de thèse vise donc à caractériser le fonctionnement morpho-sédimentaire du massif des Vosges, en cherchant à répondre à la problématique suivante : Dans quelle mesure le fonctionnement actuel et sub-actuel des rivières vosgiennes est-il impacté par les héritages (fluvio-)glaciaires et anthropiques ?

Ce travail s'articule en trois objectifs principaux, chacun correspondant à une échelle spatiale et temporelle différente. Le premier objectif est d'évaluer l'état d'équilibre sédimentaire du massif des Vosges et de mesurer les taux de dénudation des bassins versants vosgiens. Les concentrations en nucléides cosmogéniques (^{10}Be et ^{26}Al) ont été mesurées dans des échantillons de sédiments prélevés dans le lit des 21 rivières drainant l'ensemble du massif des Vosges. La mesure des concentrations en ^{10}Be - ^{26}Al a permis d'une part d'évaluer l'état d'équilibre cosmogénique de chaque bassin versant en analysant le ratio $^{26}\text{Al}/^{10}\text{Be}$, et d'autre part, de calculer les taux de dénudation pluri-millénaire dans les bassins à l'équilibre. En parallèle, la variabilité spatiale des principaux potentiels facteurs de contrôle de la dénudation a été évaluée à l'échelle du massif. A savoir, la lithologie, la morphométrie des bassins, les précipitations et les héritages glaciaires (extension des anciens glaciers et distribution des dépôts (fluvio-)glaciaires). Le deuxième objectif est de reconstituer l'évolution morpho-sédimentaire post-DMG d'un système fluvial fortement impacté par les glaciations quaternaires. Le bassin versant étudié correspond à celui de la Cleurie, qui est situé au cœur de la zone autrefois englacée et qui présente la plus forte proportion de dépôts glaciaires et fluvio-glaciaires parmi tous les bassins versants vosgiens autrefois englacés (>40% de sa surface). Le degré d'ajustement du réseau hydrographique a été évalué à partir de l'analyse des profils en long du chenal principal et de ses affluents, et de l'évolution longitudinale de la granulométrie de

surface du fond du lit. En parallèle, afin d'apporter des contraintes temporelles concernant la déglaciation de la vallée et la fermeture d'un supposé paléolac anciennement présent dans la vallée, deux profils stratigraphiques clés ont été étudiés, à partir notamment de mesures granulométriques et de datations radiocarbone. Le troisième objectif est de reconstituer la trajectoire temporelle sub-actuelle, sur les deux derniers siècles, des lits fluviaux et de leurs fonds de vallées et d'évaluer la part respective des divers facteurs de contrôle, notamment anthropiques. Les rivières concernées sont la Fecht, la Moselotte et la Vologne, qui forment un transect Est-Ouest à travers le massif des Vosges. La prospection approfondie des archives documentaires de l'administration des 'Ponts et Chaussées' a permis de construire une base de données originale, recensant l'intégralité des seuils, digues et ponts présents dans les lits et fonds de vallées des trois cours d'eau. L'évolution planimétrique diachronique des lits actifs a été reconstruite à partir de la carte d'état-major (1832) et de deux orthophotographies (1951, 2018). Enfin, à l'échelle des bassins versants, l'évolution de l'occupation du sol entre le XIX^{ème} et le XXI^{ème} siècle a été quantifiée à partir de la carte d'état-major et de l'occupation du sol actuelle.

Les résultats de cette thèse mettent en évidence une influence significative des héritages glaciaires et anthropiques sur la dynamique morpho-sédimentaire des rivières vosgiennes. À l'échelle du massif des Vosges, près de la moitié des bassins versants témoignent d'un important déséquilibre sédimentaire, corrélé à l'étendue des anciennes glaciations et des dépôts (fluvio-)glaciaires actuels. Les ratios $^{26}\text{Al}/^{10}\text{Be}$ des bassins versants en déséquilibre, compris entre 1.4 et 5.7, indiquent des temps d'enfouissement compris entre 0,5 et 3,4 Ma, suggérant des périodes de transit et de stockage des sédiments bien au-delà de l'Holocène. Le bassin versant de la Cleurie, avec un temps d'enfouissement maximum de 3,4 Ma, illustre un exemple extrême de déséquilibre sédimentaire. Les profils en long et la granulométrie de la Cleurie démontrent l'influence des héritages glaciaires sur l'incapacité du système fluvial à se rééquilibrer, impactant les processus hydro-géomorphologiques actuels et subactuels. Cette dynamique est probablement similaire dans d'autres bassins versants anciennement englacés des Vosges. L'étude des trajectoires temporelles pluri-séculaires des fonds de vallées et lits fluviaux de la Fecht, de la Moselotte et de la Vologne indique que, bien que le contexte géomorphologique spécifique à chaque hydrosystème fluvial soit le facteur de contrôle prédominant de leur dynamique morpho-sédimentaire, l'impact de la pression anthropique devient très important sur cette période. Ainsi, pour une compréhension complète des dynamiques fluviales contemporaines, il est crucial de considérer à la fois les influences glaciaires et anthropiques. Pour évaluer la capacité des cours d'eau à remobiliser les héritages glaciaires, plus de 1000 traceurs RFID ont été déployés sur 10 sites le long des trois rivières étudiées, avec des résultats attendus en 2025. Cette thèse établit un diagnostic géomorphologique de la dynamique morpho-sédimentaire des rivières vosgiennes, fournissant une base robuste pour

des décisions de gestion et de restauration des systèmes fluviaux. Ces connaissances permettront de mieux anticiper les projets de restauration hydro-géomorphologiques, de prioriser les interventions et de renforcer la résilience au changement climatique.

Abstract

Morpho-sedimentary dynamics in the Vosges Massif remain poorly understood. This gap affects both sediment flows on the current and sub-actual time scales and the long-term trajectories of sedimentary archives in the valley bottoms. Yet the long time scale is functionally dependent on the short time scale. The development of efficient and sustainable management and restoration strategies for fluvial systems must therefore take account of the long time-scale in order to better understand current morphosedimentary dynamics. The Vosges Massif has three major characteristics that make it particularly well suited to the study and comparison of the morpho-sedimentary dynamics of its river systems, using a systemic approach on nested spatio-temporal scales. First, the geological origin of its formation explains strong lithological, morphometric and climatic contrasts. Second, it has the rare characteristic of having been only partially glaciated. Finally, the Vosges Massif is the most densely populated in France, which is both a potential controlling factor of modern morpho-sedimentary dynamics, the extent of which needs to be determined, and a challenge in terms of hydrosystems management. The aim of this thesis is therefore to characterise the morpho-sedimentary functioning of the Vosges Massif, in order to address the following issue: To what extent is the current and sub-actual functioning of vosgian rivers impacted by the (fluvio-)glacial and human legacies?

This work has three main goals, each corresponding to a different spatial and temporal scale. The first goal is to evaluate the sedimentary equilibrium state of the Vosges Massif and to quantify the denudation rates of the vosgian catchments. Concentrations of cosmogenic nuclides (^{10}Be and ^{26}Al) were measured in sediment samples taken from the beds of 21 rivers draining the entire Vosges Massif. The measurement of ^{10}Be - ^{26}Al concentrations made it possible to evaluate the cosmogenic equilibrium state of each catchment by analysing the $^{26}\text{Al}/^{10}\text{Be}$ ratio, and to calculate the pluri-millennial denudation rates in the catchments at steady-state. In parallel, the spatial variability of the main potential factors controlling denudation was assessed at the massif scale. These include lithology, catchment morphometry, precipitation and glacial legacies (extension of former glaciers and distribution of (fluvio-)glacial deposits). The second goal is to reconstruct the post-LGM morpho-sedimentary evolution of a fluvial system heavily impacted by Quaternary glaciations. The catchment studied is the Cleurie, which is located at the heart of the formerly glaciated area and has the highest proportion of glacial and fluvio-glacial deposits of all the Vosges catchments that were formerly glaciated (>40% of its surface area). The current degree of adjustment of the hydrographic network was assessed by analysing the longitudinal profiles of the main channel and its tributaries, and the longitudinal evolution of the surface grain size of the riverbed. In parallel, in order to provide temporal constraints on the deglaciation of the valley and the closure of a supposed palaeolake formerly present in the valley, two key stratigraphic profiles

were studied, based on grain size measurements and radiocarbon dating. The third goal is to reconstruct the sub-actual temporal trajectory, over the last two centuries, of the riverbeds and their valley bottoms and to evaluate the respective roles of the various controlling factors, particularly anthropogenic ones. The rivers concerned are the Fecht, Moselotte and Vologne, which form an east-west transect across the Vosges Massif. An in-depth survey of the documentary archives of the 'Ponts et Chaussées' administration allowed an original database to be constructed, listing all the weirs, dykes and bridges currently present in the riverbeds and valley bottoms of the three rivers. The diachronic planimetric evolution of the active beds was reconstructed on the basis of the état-major map (1832) and two orthophotographs (1951, 2018). Finally, at catchment scale, changes in land use between the 19th and 21st centuries were quantified on the basis of the état-major map and the current land use.

The results of this thesis highlight the significant control of glacial and anthropogenic legacies on the morpho-sedimentary dynamics of vosgian rivers. At the Vosges Massif scale, almost half of the catchment are in strong cosmogenic disequilibrium, correlated with the extent of former glaciations and current (fluvio-)glacial deposits. The $^{26}\text{Al}/^{10}\text{Be}$ ratios of the catchments in disequilibrium, ranging from 1.4 to 5.7, indicate burial times of between 0.5 and 3.4 Ma, suggesting periods of sediment transit and storage well beyond the Holocene. The Cleurie catchment, with a maximum burial time of 3.4 Ma, illustrates an extreme example of sedimentary disequilibrium. The longitudinal profiles and grain size distribution of the Cleurie hydrographic network demonstrate the influence of glacial legacies on the fluvial system's inability to adjust itself, impacting current and sub-actual hydro-geomorphological processes. This dynamic is probably similar in other formerly glaciated catchments in the Vosges. The study of the multi-century temporal trajectories of the valley bottoms and riverbeds of the Fecht, Moselotte and Vologne indicates that, although the geomorphological context specific to each river hydrosystem is the predominant controlling factor in their morpho-sedimentary dynamics, the impact of human pressure is very significant over this period. For a complete understanding of contemporary fluvial dynamics, it is thus crucial to consider both glacial and anthropogenic influences. To assess the capacity of rivers to remobilise glacial materials, over 1000 RFID tracers were deployed at 10 sites along the three studied rivers, with results expected in 2025. This thesis establishes a geomorphological diagnosis of the morpho-sedimentary dynamics of vosgian rivers, providing a robust basis for decisions on the management and restoration of river systems. This knowledge will enable hydro-geomorphological restoration projects to be better anticipated, interventions to be prioritised and resilience to climate change to be strengthened.

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Abbreviations

AMS	Accelerator Mass Spectrometry
AQ	Ante-Quem
BFM	Black Forest Mountains
BRGM	Bureau de Recherches Géologiques et Minières
CEREGE	CEntre de Recherche et d'Enseignement des Géosciences de l'Environnement
CWD	Catchment-Wide Denudation
DCE	Directive Cadre sur l'Eau
DEM	Digital Elevation Model
DMG	Dernier Maximum Glaciaire
ER	Elementary Reaches
GCP	Ground Control Points
GIS	Geographical Information System
HPD	Historical Planimetric Data
HS	Hydraulic Structures
IGN	Institut National de l'Information Géographique et Forestière
INRAE	Institut National de recherche pour l'Agriculture, l'Alimentation et l'Environnement
INSEE	Institut National de la Statistique et des Etudes Economiques
LGM	Last Glacial Maximum
LIA	Little Ice Age
LVA	Longest Visible Axis
MC	Monte-Carlo
MIS	Marine Isotope Stage
MNT	Modèle Numérique de Terrain
OCSGE	Occupation du sol à Grande Echelle
OFB	Office Français de la Biodiversité

OSL	Optically Stimulated Luminescence
PAG	Petit Âge Glaciaire
PCA	Programme de Conservation et d'Actions
PQ	Post-Quem
RFID	Radio Frequency IDentification
RMSE	Root Mean Square Error
ROE	Référentiel national des Obstacles à l'écoulement
SHNEC	Société d'Histoire Naturelle et d'Ethnographie de Colmar
SVE	Spatially Variable Error
VM	Vosges Massif (or Mountains)

Chapitre 1

Introduction

1.1 Cadre thématique

1.1.1 Concepts généraux : système fluvial et hydrosystème fluvial

En géomorphologie fluviale, les concepts de système fluvial (Schumm, 1977) et d'hydrosystème fluvial (Amoros and Petts, 1993) ont été développés pour permettre l'étude systémique de la dynamique des cours d'eau, des paysages qui leur sont associés et des processus qui les façonnent.

Le système fluvial formalise les transferts de sédiments au sein du bassin versant dans une dimension principalement longitudinale, d'un point de vue théorique. Trois zones sont distinguées, dictées par l'évolution générale des pentes (Fig. 1.1). Au sein de chaque zone domine un processus différent : 1) La zone de production sédimentaire, caractérisée par de fortes pentes et dominée par l'érosion des versants. Elle fournit les têtes de bassins en sédiments. 2) La zone de transfert, caractérisée par une diminution des pentes et dominée par le transfert des sédiments vers l'aval. La remobilisation des sédiments et le dépôt de ces derniers y sont à l'équilibre. 3) La zone de dépôt, caractérisée par de faibles pentes et dominée par des processus de dépôts. Au même titre que le système fluvial dans son ensemble, ces trois zones sont des systèmes ouverts. Ils échangent de l'énergie et de la matière (eau et sédiments) avec leur environnement (Rixhon et al., 2017).

Le concept d'hydrosystème fluvial représente une extension de celui de système fluvial. Il ne se limite plus à la seule dimension longitudinale amont-aval, mais met en exergue les échanges d'énergie et de matière entre le chenal d'écoulement, sa plaine alluviale et la nappe phréatique. Ainsi, l'hydrosystème fluvial est un système à trois dimensions spatiales : longitudinale, latérale et verticale (Fig. 1.2).

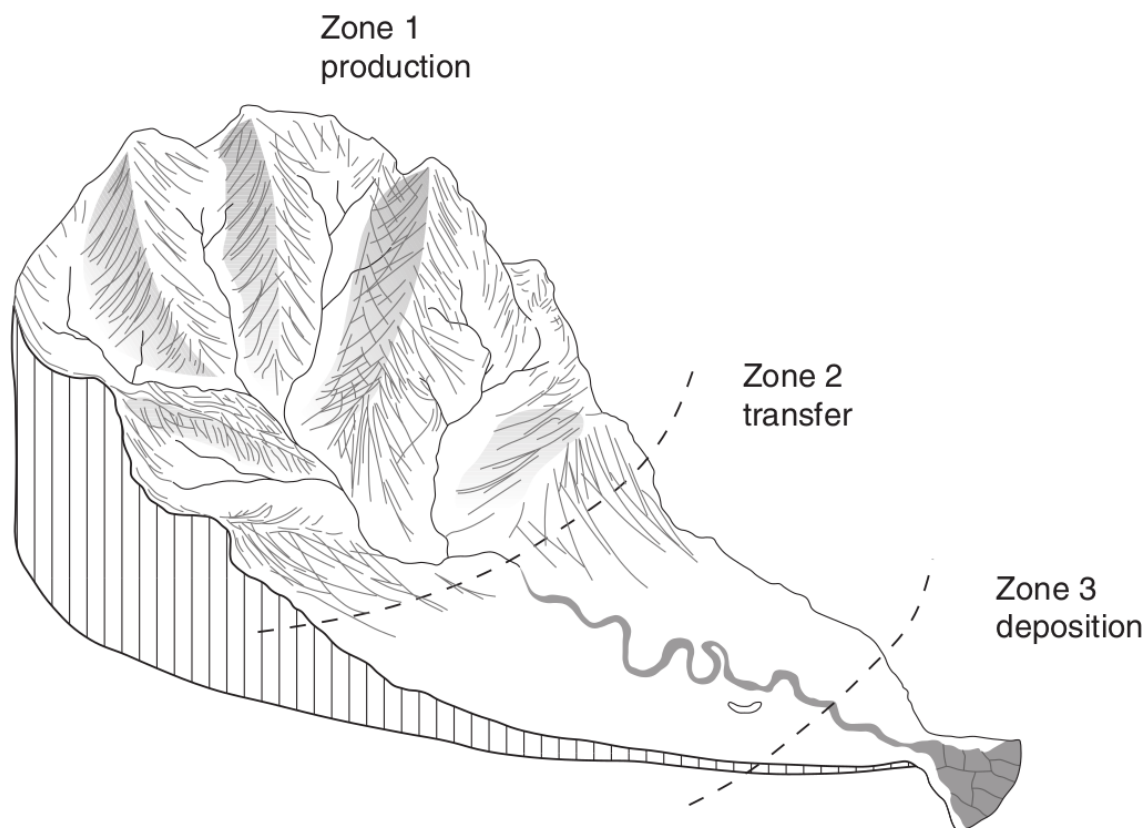


FIGURE 1.1: Représentation du concept de système fluvial (Schumm, 1977), compartimenté en trois zones. Zone 1 : production. Zone 2 : transferts. Zone 3 : dépôts. Tiré de Charlton (2008).

Les concepts de système fluvial et d'hydrosystème fluvial sont complémentaires. Le premier offre une vision globale en considérant le cours d'eau dans un ensemble, tenant compte notamment de ses interactions avec les versants et les processus hydrologiques à grande échelle (Fryirs and Brierley, 2013). Le second se concentre sur des aspects plus spécifiques, comme les interactions entre le cours d'eau, sa plaine alluviale et la nappe phréatique. Ce dernier offre ainsi une vision plus détaillée et intégrative des dynamiques hydrologiques, morpho-sédimentaires, géochimiques et écologiques, généralement dans la zone 2 du système fluvial (Fig. 1.1). En ce sens, l'hydrosystème fluvial peut être considéré comme un sous-ensemble du système fluvial. Enfin, les deux concepts reconnaissent l'importance de la dimension temporelle dans l'évolution des cours d'eaux (Schumm, 1977; Amoros and Petts, 1993). Ils reconnaissent ces derniers comme des systèmes dynamiques qui évoluent dans le temps et l'espace en réponse à des facteurs environnementaux et humains.

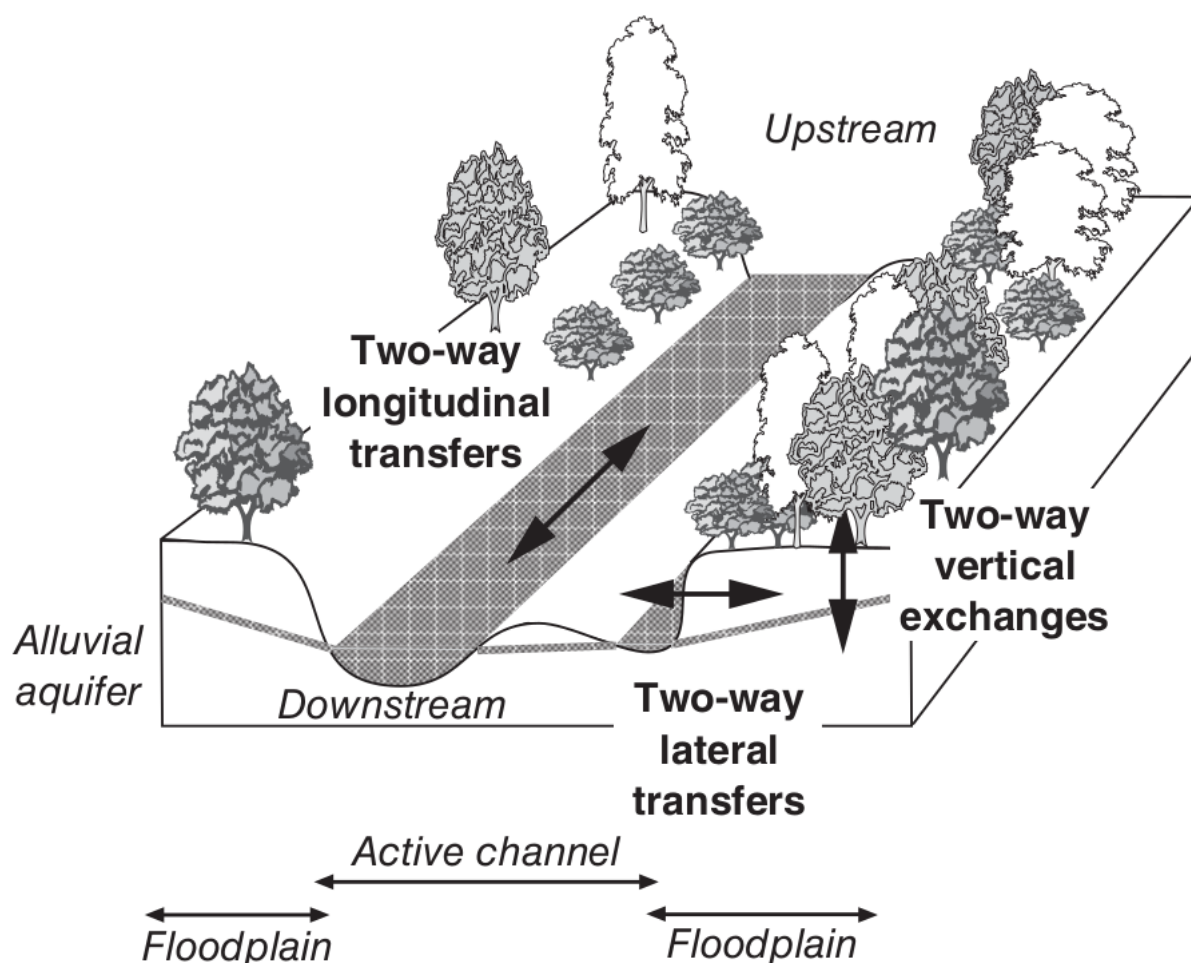


FIGURE 1.2: Représentation du concept d'hydrosystème fluvial (Amoros and Petts, 1993) et de ses trois dimensions spatiales. Transferts longitudinaux au sein du chenal actif. Transferts latéraux entre le chenal actif et la plaine inondable. Transferts verticaux avec la nappe phréatique. Tiré de Piégay (2016).

1.1.2 La dynamique morpho-sédimentaire d'un cours d'eau : variables de contrôle et d'ajustement, équilibre et perturbations

La dynamique temporelle d'un système fluvial détermine un fonctionnement de type processus-réponse au sein du bassin versant, entre les cascades sédimentaires amont-aval et les morphologies fluviales associées aux différentes zones du système (Piégay, 2016). Ainsi, les dynamiques spatiales et temporelles du système fluvial conditionnent la dynamique morpho-sédimentaire du cours d'eau, caractérisée par ses variations morphologiques et sédimentaires dans le temps et l'espace. A l'échelle du bassin versant, la dynamique du système fluvial est contrôlée par des variables dites externes (indépendantes) : le climat, la tectonique, le niveau de base et l'activité humaine (Schumm, 1977, Fig. 1.3). Ces variables opèrent indépendamment et ne sont pas influencées par le fonctionnement interne du système fluvial. En réponse à ce contrôle externe, des ajustements s'opèrent au sein du bassin versant, à travers des variables dites internes (dépendantes). Les deux principales sont le débit liquide (Ql) et le débit

solide (Q_s) du cours d'eau. D'autres variables internes correspondent à des paramètres morphométriques (ex. pente des versants, densité de drainage) et morphologiques (ex. largeur-profondeur du chenal, style fluvial, granulométrie du lit, etc.) ou encore au type de sol ou à la couverture végétale des versants (Fig. 1.3).

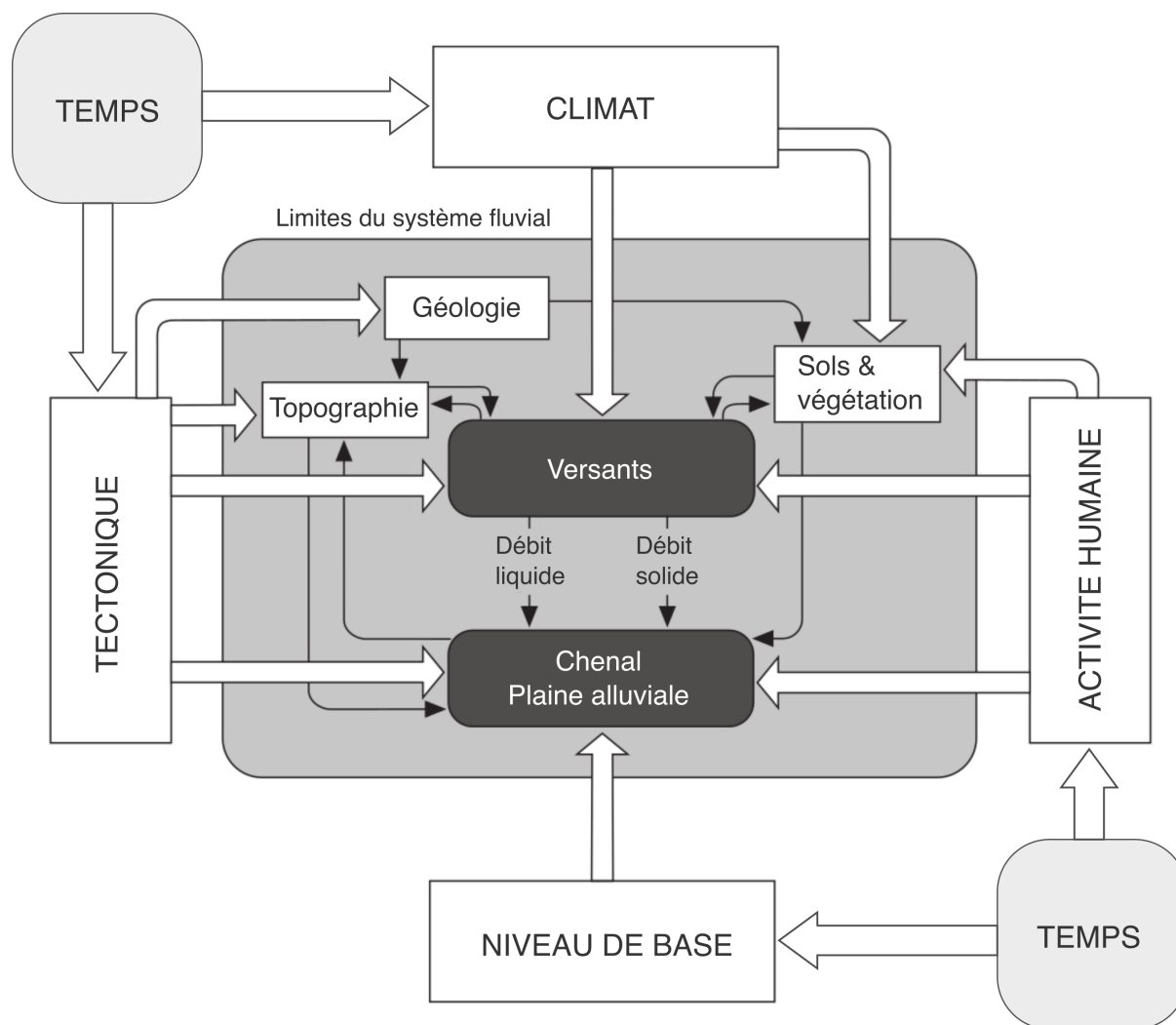


FIGURE 1.3: Représentation schématique simplifiée du système fluvial et de ses variables de contrôles externes et internes. La variable temporelle y a été ajoutée comme variable de contrôle externe. Tiré et modifié de Charlton (2008).

A l'échelle d'un sous-système cependant, comme un tronçon de cours d'eau, Q_l et Q_s peuvent être considérées comme des variables externes, car les apports d'eau et de sédiments dans ce tronçon particulier proviennent de l'amont du bassin versant et contrôlent les ajustements morphodynamiques de ce tronçon (Charlton, 2008). De manière générale, les variables internes d'ajustement sont donc influencées par les variables externes de contrôle, mais elles peuvent l'être également par d'autres variables internes, selon l'échelle spatiale d'étude adoptée. Le débit solide peut par exemple être influencé par la couverture végétale, elle-même contrôlée par le climat et l'activité humaine. Enfin, le temps lui-même doit être considéré à

la fois comme une variable de contrôle externe du système fluvial (Fig. 1.3) et comme la quatrième dimension de ce dernier (Piégay, 2016). En effet, chaque système fluvial présente des héritages résultant de son histoire géologique, climatique et anthropique, qui conditionnent la dynamique morpho-sédimentaire actuelle du cours d'eau (Lespez et al., 2015).

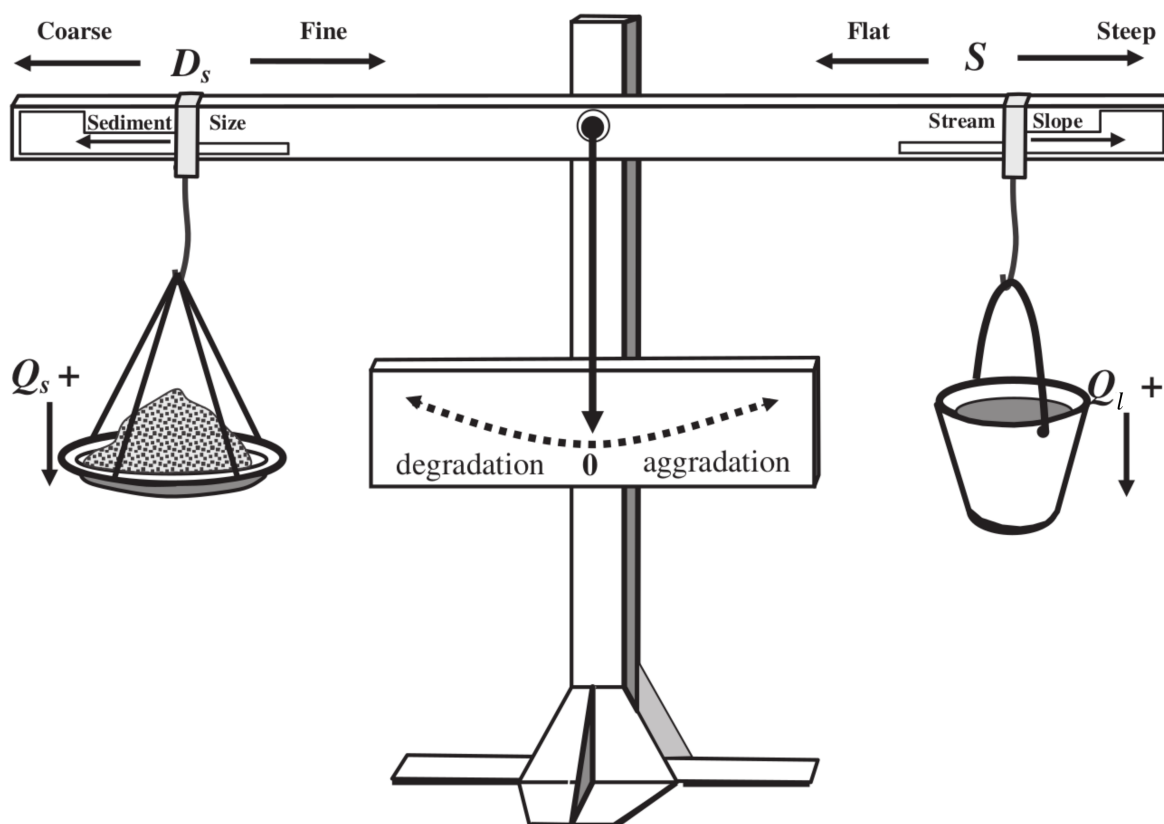


FIGURE 1.4: Représentation du concept de la 'Balance de Lane' (Lane, 1955) et du principe d'équilibre dynamique. Tiré de Dust and Wohl (2012).

En théorie, la dynamique morpho-sédimentaire d'un cours d'eau est à l'état d'équilibre dynamique (Church and Ferguson, 2015). Cela signifie que ses variables d'ajustements (Q_l et Q_s) oscillent autour d'une valeur moyenne, en réponse aux oscillations des variables de contrôles, qu'elles soient externes ou internes. Cet état d'équilibre dynamique a été formalisé par Lane (1955) puis illustré par Borland (1960) à travers la 'Balance de Lane' (Fig. 1.4). L'état d'équilibre dynamique d'un cours d'eau implique un ajustement permanent de ce dernier, en réponse aux variations de Q_l et Q_s , qui surviennent aussi bien à l'échelle annuelle (saisons hydrologiques) que sur le temps plus long (cycles glaciaires ou événements climatiques majeurs). L'ajustement du cours d'eau se traduit par des changements morphologiques, à travers notamment des processus d'aggradation/incision de son lit, ou encore d'érosion de ses berges. Ces ajustements contribuent au maintien de son état d'équilibre dynamique, définissant son état moyen, avec un certain degré de stabilité (Knighton, 1984). Cette auto-régulation continue du système fluvial, alimentée par des rétroactions, favorise sa résilience en lui permettant de revenir à un nouvel équilibre après une perturbation (Bravard and Petit, 1997).

Lorsqu'une perturbation survient, qu'elle soit d'origine climatique ou anthropique, le système réagit en fonction de l'ampleur et de la durée de la perturbation, ainsi que de ses capacités de résilience (Fig. 1.5). Une perturbation forte mais de courte durée peut conduire le cours d'eau vers un nouvel état, suivi d'un retour à son état initial après un certain temps de réaction. En revanche, des modifications de longue durée et/ou des perturbations de magnitude particulièrement élevée, peuvent entraîner des fluctuations plus significatives vers un nouvel état d'équilibre pouvant se maintenir sur une longue période, qui peut alors correspondre à une 'métamorphose fluviale' (Bravard and Petit, 1997; Allée, 2003). Enfin, les ajustements d'un cours d'eau aux forçages environnementaux ou anthropiques varient en fonction de l'échelle spatiale des formes (Fig. 1.6), influençant ainsi le temps de réaction du chenal aux modifications des variables de contrôle (Fig. 1.5). A l'échelle du bassin versant par exemple, la pente du profil en long d'un cours d'eau évolue à l'échelle pluri-millénaire, tandis qu'un changement de la forme en plan d'un hydrosystème fluvial peut s'effectuer en quelques années, voire en un seul événement hydrologique (Fig. 1.6).

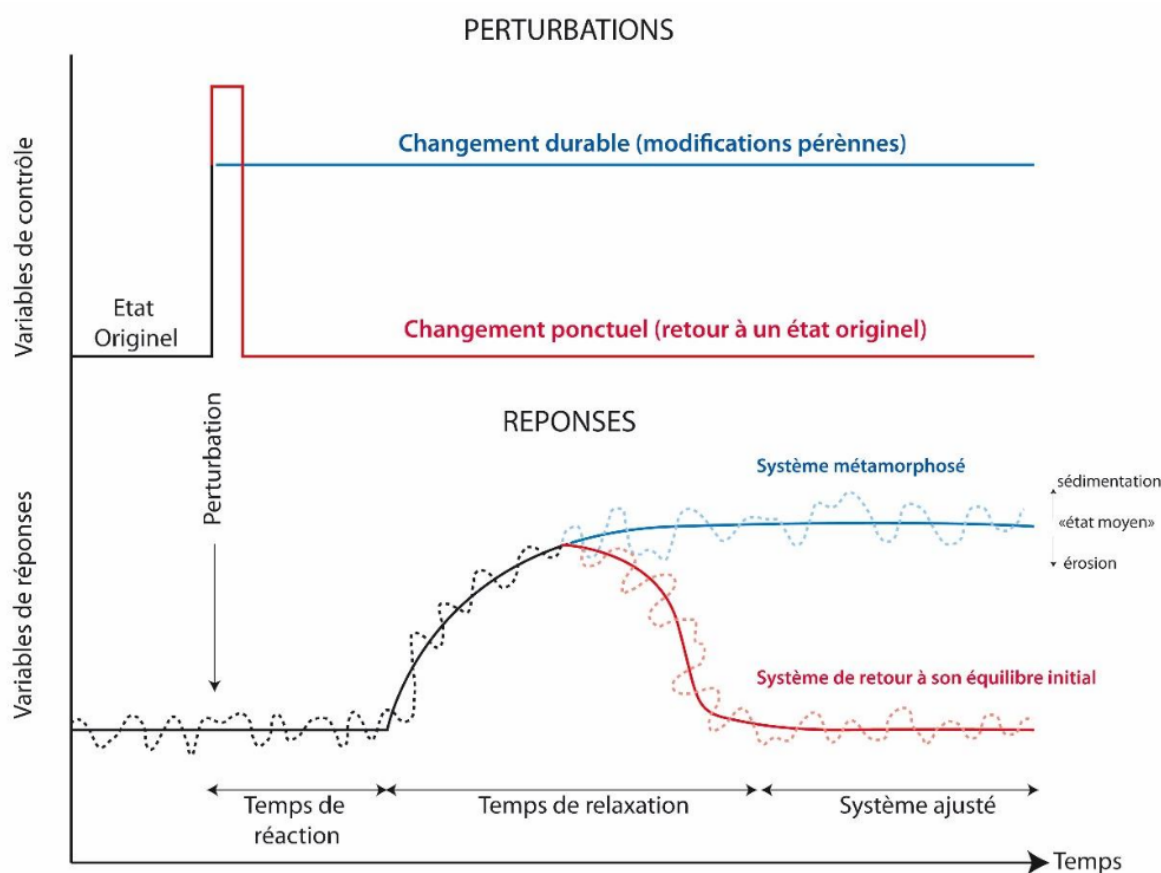


FIGURE 1.5: Équilibre dynamique et réponse théorique d'un système fluvial à différents types de perturbations (Knighton, 1984). Adapté par Melun (2012). Tiré de Beauchamp (2018).

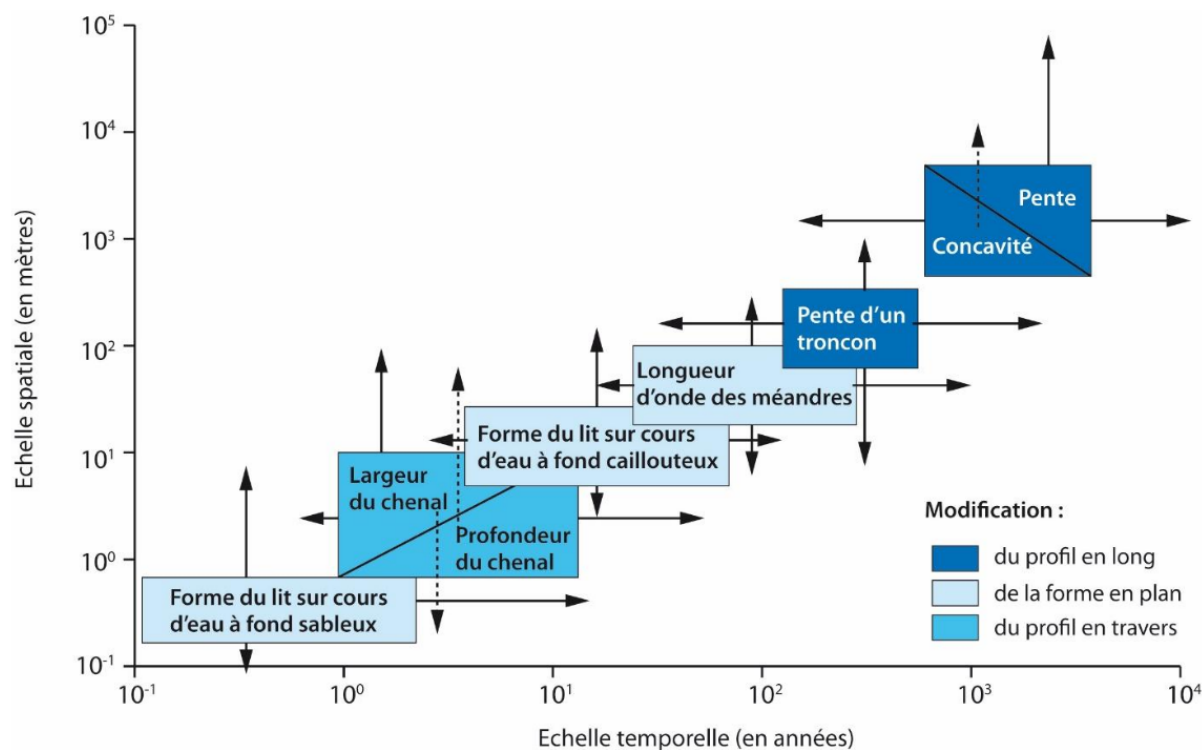


FIGURE 1.6: Emboîtement des échelles spatiales et temporelles des ajustements morphologiques au sein d'un système fluvial (Knighton, 1984). Tiré de Beauchamp (2018).

1.1.3 Diagnostic géomorphologique et restauration d'un système fluvial : importance de l'emboîtement des échelles spatio-temporelles

Le fonctionnement actuel d'un cours d'eau est donc le fruit d'une combinaison complexe de facteurs de contrôle externes et internes au système fluvial (Fig. 1.3), qui opèrent à des échelles spatio-temporelles emboîtées (Fig. 1.6). Élaborer un diagnostic géomorphologique d'un système fluvial revient donc à adopter une approche systémique, visant à étudier les relations fonctionnelles entre les différents compartiments du système fluvial (Fig. 1.1) et leurs interactions (Lespez, 2007, ; Fig. 1.3). Cette approche permet notamment de tenir compte des différents héritages, d'origine climatique, tectonique ou anthropique, qui conditionnent la dynamique morpho-sédimentaire actuelle du cours d'eau. Bien qu'encore rarement adoptée, cette approche est de plus en plus préconisée pour permettre la compréhension du fonctionnement actuel des systèmes fluviaux (Allée, 2003; Brierley and Fryirs, 2013; Scapozza and Oppizzi, 2013). Elle consiste à appréhender les processus de surface et leur modalités, comme la dénudation des versants et le transport des sédiments dans les corridors fluviaux, d'une part sur le temps long (dizaines de milliers d'années) et d'autre part sur des échelles temporelles plus courtes (pluri-séculaire, pluri-décennale voire pluri-annuelle). Dans le premier cas, les facteurs qui contrôlent la formation et l'évolution des paysages sont principalement d'origine tectonique et climatique. Dans le deuxième cas, l'influence humaine historique sur les paysages joue un rôle prédominant dans la dynamique morpho-sédimentaire des cours

d'eau, bien que le signal climatique puisse également contribuer, notamment à travers des événements comme le Petit Âge Glaciaire (PAG), sachant que les effets de ce dernier peuvent à leur tour être amplifiés par l'impact de l'activité humaine sur les bassins versants (Lespez et al., 2015; Delile et al., 2016; Brown et al., 2018).

L'approche systémique est également d'autant plus pertinente dans un contexte de mise en place de stratégies de restauration de cours d'eau, encadrées en Europe par la Directive Cadre sur l'Eau (DCE) adoptée en 2000. Or, si l'un des objectifs de la DCE est le retour à un 'bon état écologique' des cours d'eau (Schmitt et al., 2007; Gob et al., 2014; Downs and Piégay, 2019), la pertinence d'une restauration vers un état 'naturel' est encore largement débattue, car il est difficile à identifier et n'est pas nécessairement réalisable, ni même souhaitable, en fonction de l'occupation et des besoins humains actuels (Brown et al., 2018). De plus l'état 'naturel' d'un cours d'eau est discutable, considérant la nature 'hybride' des hydrosystèmes, fruit des interactions complexes entre les processus naturels et les activités humaines (Lespez and Dufour, 2021). Il peut alors être souhaitable de viser un état qui corresponde à un état morphologique du cours d'eau actuellement 'atteignable' (Brierley and Fryirs, 2013). Cet état doit tenir compte du fonctionnement actuel du cours d'eau, conditionné par son histoire et ses héritages, qu'ils soient climatiques et/ou anthropiques. Ainsi, l'élaboration d'un diagnostic géomorphologique d'un système fluvial est une étape clé dans la détermination d'objectifs de restauration atteignables et souhaitables. Malheureusement, cette approche n'est pratiquement jamais intégrée par les gestionnaires des cours d'eau (Bravard, 2002; Schmitt, 2011; Mason and Polvi, 2023). Elle peut pourtant permettre de répondre à des questions cruciales telles que : Les cours d'eau sont-ils en cours d'ajustement ? Sont-ils déjà ajustés à des conditions spécifiques ? Quelle est leur capacité à se réajuster ? Ont-ils subi une métamorphose fluviale, et si oui, depuis quand et quels en ont été les facteurs de contrôle ?

Dans cette perspective, l'étude de la dynamique morpho-sédimentaire des rivières vosgiennes (massif des Vosges, Nord-Est de la France) en vue de l'élaboration d'un diagnostic géomorphologique fonctionnel semble particulièrement pertinente, tant d'un point de vue fondamental qu'appliqué. Premièrement, la dynamique morpho-sédimentaire dans le massif des Vosges reste aujourd'hui mal connue. Cette lacune touche autant les flux sédimentaires aux échelles temporelles actuelle et subactuelle que les trajectoires temporelles longues des stocks sédimentaires des fonds de vallées postérieurs au dernier maximum glaciaire (Cordier et al., 2017). Deuxièmement, l'origine géologique du massif des Vosges explique un fort contraste Nord-Sud et Est-Ouest de ses caractéristiques lithologiques, morphométriques et climatiques, ce à quoi s'ajoute le fait que le massif n'ait été englacé que partiellement (Flageollet, 2002). Ces spécificités font du massif des Vosges un cadre sans doute unique en France pour tenter de déconvoluer la part respective des différents facteurs de contrôle de la dynamique morpho-sédimentaire actuelle des rivières vosgiennes. Enfin, le massif des Vosges est le plus densément peuplé de France et présente de nombreuses structures hydrauliques, vestiges de

l'industrialisation de ses vallées (Parmentier, 2019). Cela constitue à la fois un facteur de contrôle potentiel de la dynamique morpho-sédimentaire actuelle des rivières vosgiennes, dont il convient de déterminer l'ampleur, et un enjeu en termes de gestion des hydrosystèmes fluviaux.

1.2 Zone d'étude

1.2.1 Le massif des Vosges

Le massif des Vosges est un massif de moyenne montagne d'origine hercynienne situé au Nord-Est de la France (Fig. 1.7). Il forme le horst occidental du système Vosges-Palatinat/Fossé Rhénan/Forêt Noire-Odenwald, qui s'est élevé au Miocène lors d'un plissement lithosphérique, conséquence de contraintes de compression dans l'avant-pays alpin, tandis que l'effondrement du Fossé Rhénan est attribué à une extension lithosphérique (Ziegler and Dèzes, 2007).

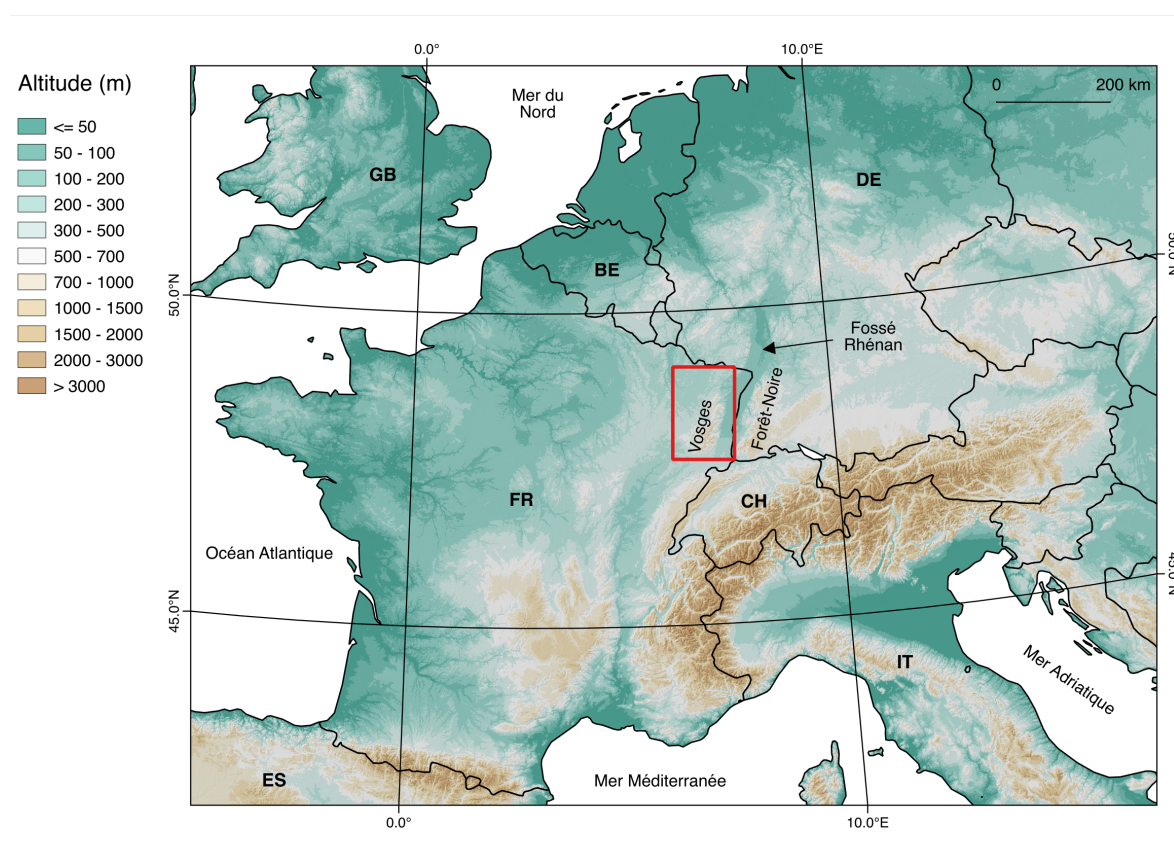


FIGURE 1.7: Localisation du massif des Vosges en Europe Occidentale. L'emprise du rectangle rouge correspond à l'agrandissement présenté dans les figures 1.8 et 1.9.

Le massif des Vosges est orienté NNE/SSO (Fig. 1.8A). Il a une longueur d'environ 180 km et une largeur variant de 60 km (Epinal-Colmar) à 10 km (col de Saverne). En termes de topographie, le massif présente un double gradient, N-S et E-O. D'une part, l'altitude moyenne

du massif varie d'environ 300 m dans la partie Nord à plus de 1200 m dans la partie Sud (Fig. 1.8A). D'autre part, le versant Est présente des pentes abruptes vers le Fossé Rhénan, tandis que les pentes du versant Ouest sont plus douces vers le Plateau Lorrain, qui constitue la partie orientale du Bassin Parisien. Une vingtaine de bassins versants principaux drainent le massif. Ceux à l'Est confluent pour la majorité vers l'Ill, elle-même affluent du Rhin (Fig. 1.8A). Sur le même versant, les cours d'eau au Nord de la Bruche confluent directement avec le Rhin. Le versant Ouest des Vosges est drainé par la Meurthe et la Moselle et leurs principaux affluents (Fig. 1.8A). La lithologie des Vosges est grossièrement bipartite : les roches cristallines paléozoïques et les grès triasiques composent globalement les parties sud ($\sim 2/3$) et nord ($\sim 1/3$) du massif (Fig. 1.8B). Contrastant fortement avec la couverture sédimentaire mésozoïque presque uniforme au nord, le socle paléozoïque présente des lithologies variées (Skrzypek, 2011). Les roches plutoniques prédominent (granites et (grano-)diorites principalement), mais des roches métamorphiques (surtout gneiss) et volcaniques (laves et tufs) sont également présentes. Le massif des Vosges est nettement délimité à l'est par les systèmes de failles normales Vosgien et Rhénan, qui borde le Fossé Rhénan, tandis que le versant Ouest tombe sur les formations mésozoïques légèrement inclinées du Plateau Lorrain (Fig. 1.8B).

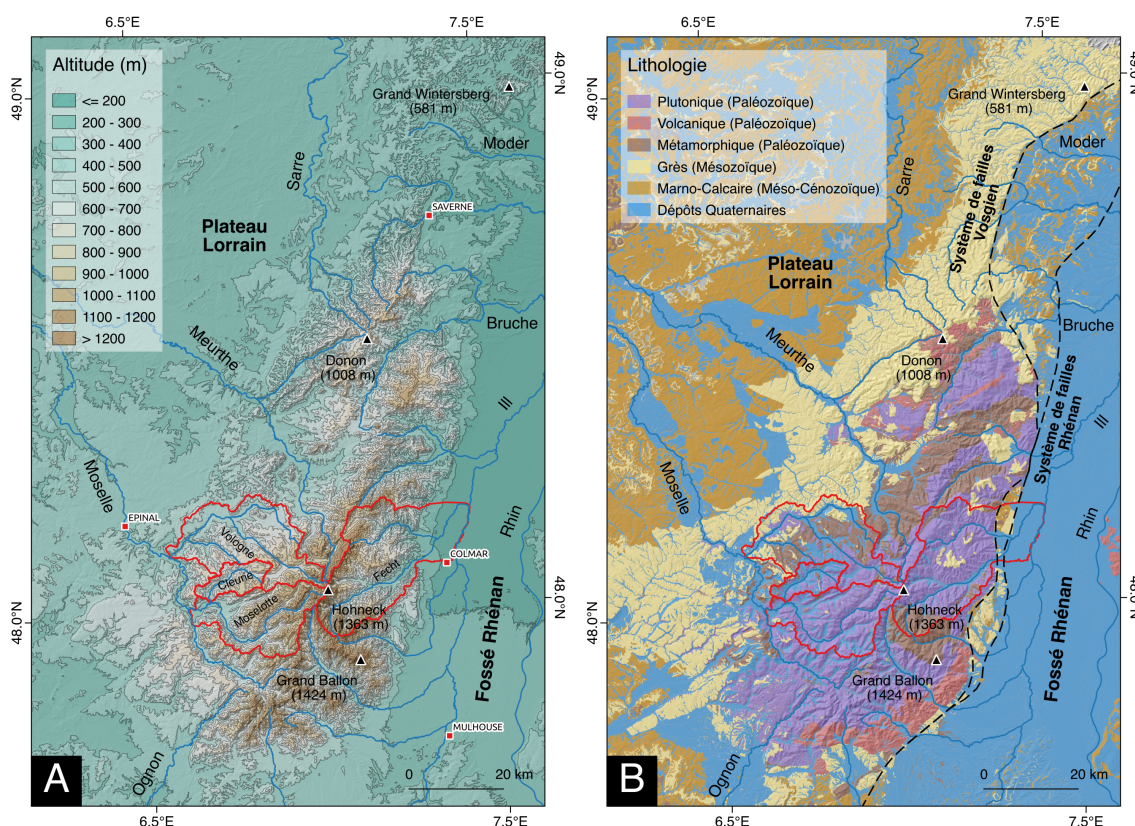


FIGURE 1.8: Carte (A) topographique et (B) géologique du massif des Vosges et principaux cours d'eau drainés par l'Ill (versant Est) et la Moselle et la Meurthe (versant Ouest). Source : Carte géologique harmonisée du BRGM au 1/50 000. Les bassins versants de la Fecht, Moselotte, Vologne et Cleurie sont mis en évidence en contours rouges.

La distribution spatiale des glaciations quaternaires et leurs empreintes dans le paysage reflètent également une forte dichotomie N-S (Fig. 1.9A-B ; Flageollet, 2002; Mercier and Jeser, 2004; Mercier, 2014). Le tiers septentrional du massif se caractérise par des paysages exclusivement fluviaux. Ils ne présentent aucune trace d'activité glaciaire antérieure et permettent de déduire une zone libre de glace dans le tiers nord du massif (Flageollet, 2002; Mercier, 2014). Les paysages fluviaux de la partie centrale témoignent localement et de manière éparse d'une activité glaciaire passée de faible magnitude, c'est-à-dire de cirques et de moraines de cirque de petite taille (Darmois-Théobald, 1972; Valenti et al., 2024). En revanche, le tiers sud du massif présente des paysages nettement marqués par les glaciations. De nombreux cirques bien développés de part et d'autre de la principale ligne de crête, des vallées en auge, des dépôts glaciaires et fluvio-glaciaires recouvrant les fonds de vallées (Fig. 1.9B) indiquent tous une forte activité glaciaire passée. Malgré la quasi-absence de datation numérique, la partie sud du massif des Vosges témoigne de plusieurs glaciations quaternaires caractérisées par diverses emprises spatiales (Fig. 1.9A). Flageollet (2002) reconnaît au moins trois à cinq glaciations. L'englacement passé de la partie sud du massif des Vosges est également bien illustré par la présence de dépôts glaciaires et fluvio-glaciaires, absents au nord (Fig. 1.9B). Enfin, les sondages géologiques du Bureau de Recherches Géologiques et Minières (BRGM) mettent en évidence un gradient N-S marqué en termes d'épaisseurs minimums de sédiments (non-différenciés) présents dans les fonds de vallées (Fig. 1.9B), entre la partie nord non-englacée et la partie sud plusieurs fois englacée.

En raison de l'orientation NNE/SSO de sa principale ligne de partage topographique, le massif des Vosges forme une barrière orogénique presque perpendiculaire au flux atmosphérique occidental océanique prédominant, induisant ainsi un effet de Foehn marqué (Fig. 1.9C; François and Humbert, 2000). Les précipitations annuelles moyennes augmentent également avec l'altitude le long de la crête principale, allant d'environ 800 mm/an dans la partie nord jusqu'à 1400 mm/an à des altitudes supérieures à 1200 m (Fig. 1.9C).

Concernant l'occupation actuelle du sol, le massif des Vosges est très largement forestier (Fig. 1.9D). Plus de 70% de sa surface est recouverte par des forêts de feuillus et de résineux, ce qui en fait, à l'échelle de la France, le massif montagneux le plus densément boisé (Rochel, 2004). Avec une densité de population d'environ 100 hab/km² (INSEE), le massif des Vosges est également considéré comme l'un des plus densément peuplés d'Europe (Parmentier, 2019). Cette anthropisation a été progressive et remonte à plusieurs millénaires (Goepp, 2007; Nüsslein, 2016). Elle peut s'expliquer à la fois par la localisation du massif comme carrefour et frontière à l'échelle de l'Europe, mais également par la présence de divers minerais et de nombreux cours d'eau permettant l'irrigation agricole des fonds de vallées et la production d'énergie hydraulique (Parmentier, 2019). Durant l'Antiquité, le massif des Vosges est habité par des populations gallo-romaines qui pratiquent l'agriculture, l'élevage et l'exploitation forestière (Toutain, 1940; Schwartz et al., 2005; Gouriveau et al., 2020). Au cours du Moyen

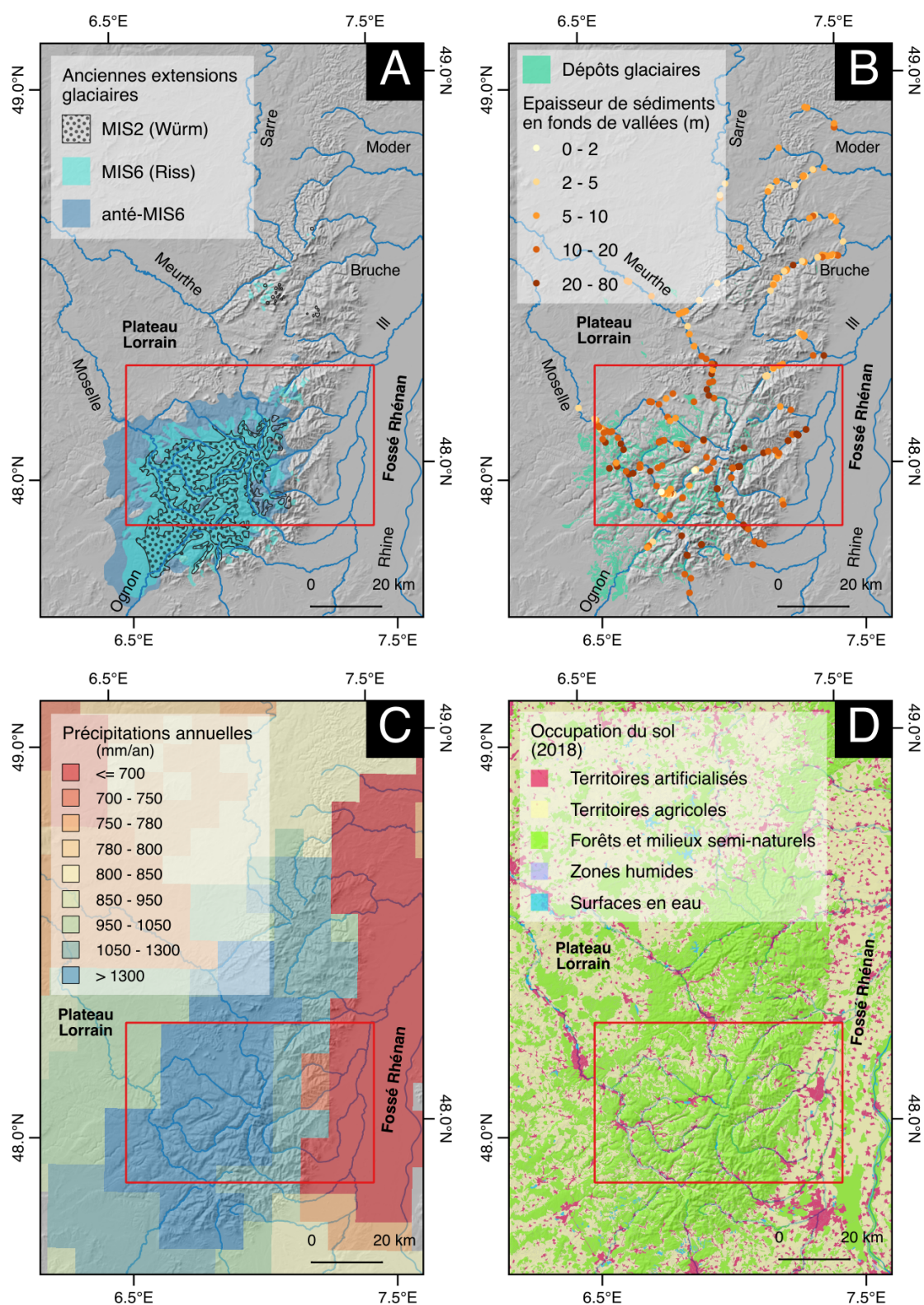


FIGURE 1.9: Cartographie des principaux contrastes climatiques quaternaires et actuels à l'échelle du massif des Vosges. Distribution spatiale (A) des extensions glaciaires quaternaires d'après Flageollet (2002), (B) des dépôts glaciaires et fluvio-glaciaires, ainsi que de l'épaisseur des dépôts de fonds de vallées, d'après la carte géologique du BRGM, (C) des précipitations annuelles moyennes, d'après les séries climatiques Météo-France (Vidal et al., 2010) sur la période 1958-2010 et (D) de l'occupation actuelle du sol d'après OCSGE (2018). L'emprise du rectangle rouge correspond à l'agrandissement présenté dans les figures 1.10 et 1.11.

Âge, l'exploitation forestière s'est largement développée pour répondre aux besoins croissants en bois de construction, de chauffage et pour la fabrication de charbon de bois (Keller, 2023). L'exploitation minière est devenue importante à partir de la période médiévale tardive, plus particulièrement à la Renaissance, avec l'extraction de minerais tels que le fer, le cuivre et le plomb (Fluck and Ancel, 1989). À partir de la révolution industrielle (milieu du XIX^{ème} siècle), le massif des Vosges est le théâtre d'une industrialisation rapide, avec le développement de l'industrie textile, minière et métallurgique (Parmentier, 2019). Cette industrialisation est portée par une exploitation forestière massive, constituant la dernière phase de défrichement du massif (Gebhardt et al., 2015). Aujourd'hui, bien que les activités industrielles aient considérablement décliné, les fonds de vallées vosgiens sont toujours très densément urbanisés (Parmentier, 2019, Fig. 1.9D).

1.2.2 Les bassins versants de la Fecht, Moselotte, Vologne et Cleurie

Cette étude se focalise dans un second temps sur quatre bassins versants adjacents, situés dans la partie sud du massif ; il s'agit de ceux de la Fecht, de la Moselotte, de la Vologne et de la Cleurie (Fig. 1.8A). Ils forment un transect Est-Ouest de part et d'autre de la ligne de crête du massif, au sein duquel nous voulons comparer les dynamiques morpho-sédimentaires de leur cours d'eau et les facteurs de contrôle correspondants. Leurs principales caractéristiques morphométriques, lithologiques et d'englacement passé sont présentées ci-dessous (Figs. 1.10-1.11).

La Fecht, affluent de l'Ill, a une longueur de 49 km et une aire drainée de 520 km². Son fond de vallée est particulièrement large, notamment à partir du deuxième tiers aval, vers la ville de Munster (largeur ~1000 m; Fig. 1.10). La Fecht a formé un large cône de déjection dans le Fossé Rhéna, à la sortie du massif (Fig. 1.10). La lithologie du bassin de la Fecht est principalement plutonique, avec des affleurements métamorphiques en amont et des reliquats de grès sur l'interfluve nord (Fig. 1.10). Lors des dernières glaciations, plusieurs langues glaciaires ont probablement façonné la partie amont de la vallée, potentiellement jusqu'à Munster (Fig. 1.11; Andréoli et al., 2006; Flageollet, 2002). En tête de bassin uniquement, des dépôts (fluvio-)glaciaires sont cartographiés dans le fond de vallée et sur quelques versants. Une série de cirques abrupts et de petits lacs de surcreusement glaciaires (lac du Schiessrothried, lac d'Altenweiher) sont présents le long du flanc oriental. Les sondages géologiques effectués dans le fond de vallée indiquent des épaisseurs de sédiments variant de 13 à 37 m (Fig. 1.10). L'épaisseur atteint 53 m au niveau de l'apex du cône de déjection.

La Moselotte, affluent de la Moselle, a une longueur de 47 km et une aire drainée de 361 km². Son fond de vallée est fortement encaissé sur les deux tiers amont (largeur ~100 m), puis s'élargit significativement dans le dernier tiers (largeur ~1000 m; Fig. 1.10). Sur les

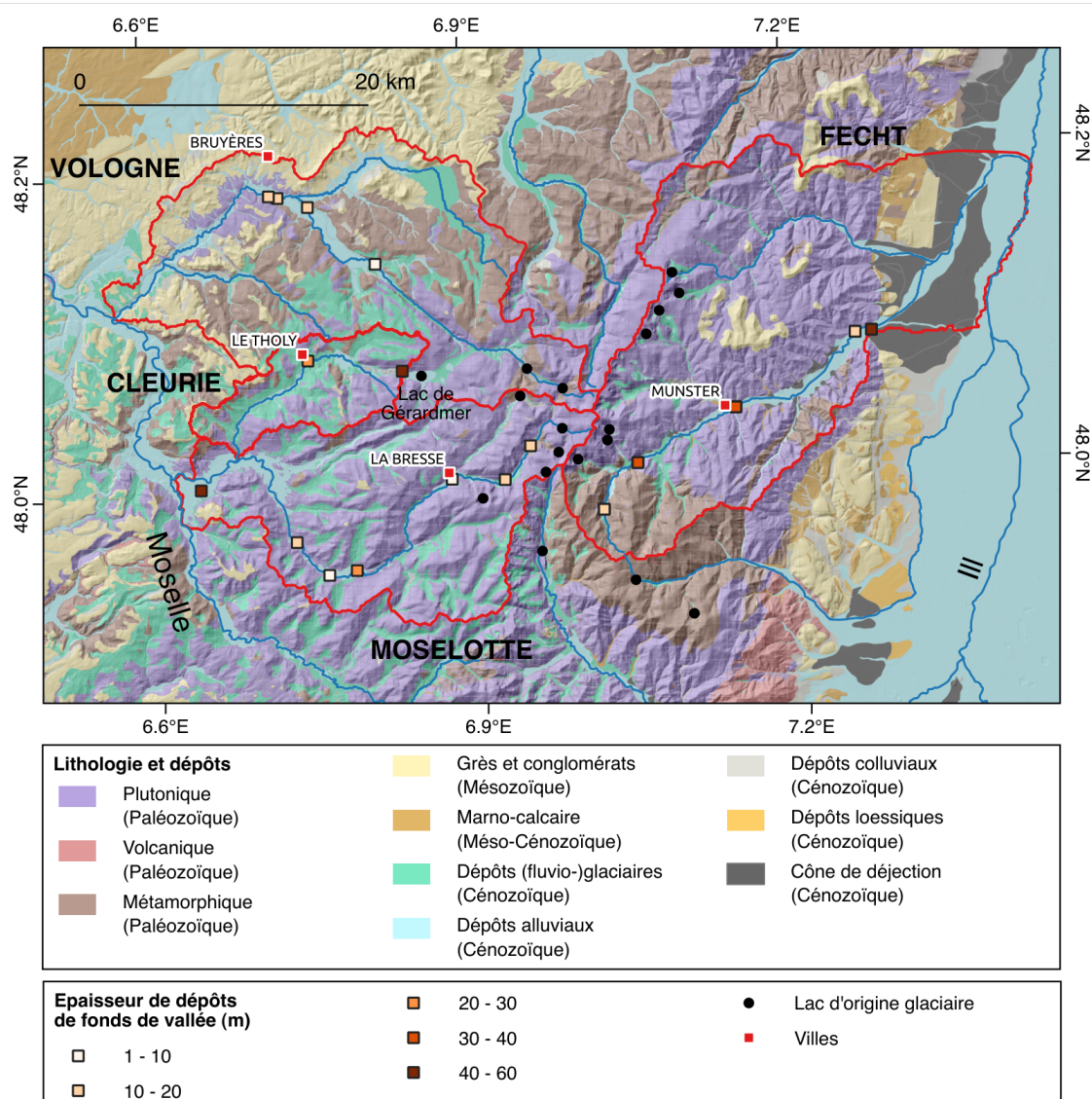


FIGURE 1.10: Lithologie et dépôts quaternaires (glaciaires, fluvio-glaciaires et alluviaux) au sein des bassins versants de la Fecht, Moselotte, Vologne et Cleurie. Source : carte géologique harmonisée (BRGM).

deux derniers kilomètres, la Moselotte termine son cours dans la vallée de la Moselle, qui correspond à un paléolac (Cordier, 2015). La lithologie du bassin de la Moselotte est très largement plutonique (Fig. 1.10). La glaciation rissienne aurait recouvert l'intégralité du bassin versant de la Moselotte, tandis que la glaciation würmienne aurait été caractérisée par une série de langues glaciaires massives (Fig. 1.11). Une remarquable difffluence glaciaire est à noter à la confluence avec la Cleurie (Seret, 1967; Flageollet, 2002). Des dépôts (fluvio-)glaciaires sont cartographiés de manière généralisée dans le bassin versant, ainsi que sur une partie du fond de vallée, particulièrement dans la moitié amont (Fig. 1.10). Quelques lacs de surcreusement glaciaires sont à noter en tête de bassin, dont le lac de la Lande, traversé par la Moselotte. Les sondages géologiques effectués dans le fond de vallée indiquent des épaisseurs

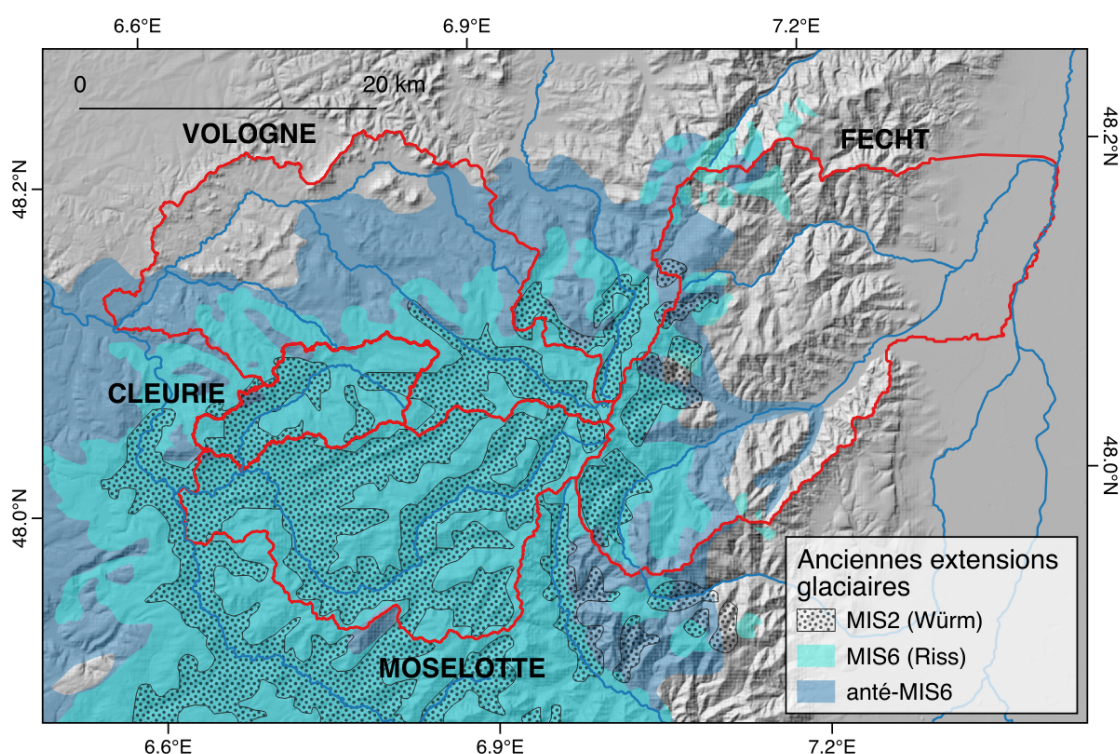


FIGURE 1.11: Carte de l'extension glaciaire supposée au MIS2, MIS6 et des glaciations antérieures au MIS6 sur les bassins versants de la Fecht, Moselotte, Vologne et Cleurie, d'après Flageollet (2002).

de sédiments variant de 10 à 30 m (Fig. 1.10). L'épaisseur atteint 54 m à la confluence avec la Moselle.

La Vologne, affluent de la Moselle, a une longueur de 50 km et une aire drainée de 367 km². Sa vallée est fortement encaissée sur le premier tiers amont (largeur ~100 m), elle s'élargit ensuite considérablement sur le deuxième tiers (largeur ~1000 m) et finit par se contracter à nouveau (largeur ~ 500m) avant sa confluence avec la Moselle, aux alentours de la ville de Bruyères (Fig. 1.10). La lithologie du bassin de la Vologne est relativement variée. Elle présente des roches essentiellement plutoniques en amont, métamorphiques dans la partie médiane et principalement gréseuses en aval (Fig. 1.10). Les glaciations n'auraient occupé que partiellement le bassin de la Vologne (Seret, 1967; Flageollet, 2002). Au Würm, deux langues glaciaires se seraient développées d'une part dans la gorge de Kichompré au Nord-Ouest, et d'autre part vers l'actuelle tête de bassin de la Cleurie à l'Ouest (Fig. 1.11). Les glaciations antérieures ne se seraient étendues guère plus au Nord que l'actuelle ville de Bruyères. Les dépôts glaciaires cartographiés tapissent certains versants et s'observent également dans la vallée, exclusivement dans la moitié amont du bassin. Le plus grand lac glaciaire du massif des Vosges, le lac de Gérardmer (surface ~1,16 km²) est localisé en amont du bassin de la Vologne. Sa moraine terminale correspond à l'actuel interfluve entre le bassin de la Vologne

et celui de la Cleurie. Les sondages géologiques effectués dans le fond de vallée indiquent des épaisseurs de sédiments variant de 9 à 17 m (Fig. 1.10).

La Cleurie, affluent de la Moselotte, a une longueur de 19 km. Son bassin versant est adjacent à celui de la Moselotte, de la Vologne et de la Moselle. Il draine une surface de 78 km². Sa vallée est particulièrement large en tête de bassin (largeur ~600 m). Elle se rétrécit assez brusquement à partir de la ville du Tholy (largeur ~250 m; Fig. 1.10). La lithologie du bassin de la Cleurie est majoritairement plutonique (Fig. 1.10). L'histoire glaciaire du bassin de la Cleurie est singulière. Au Würm, trois langues glaciaires y auraient convergé, dont la diffuence du glacier de la Moselotte, provoquant l'apparition d'un lac au centre de la vallée (Flageollet, 1988, 2002). Le bassin de la Cleurie est recouvert à quasiment 50 % de dépôts (fluvio-)glaciaires, localisés sur les versants et dans le fond de vallée, dont une série de moraines terminales et des dépôts deltaïques bien conservés. Les deux sondages géologiques effectués dans le fond de vallée indiquent des épaisseurs de sédiments variant de 25 et 54 m.

1.3 Problématisation, objectifs et méthodes

Ce travail de thèse vise à apporter des nouvelles connaissances, fondamentales et appliquées, sur le fonctionnement morpho-sédimentaire du massif des Vosges. A travers l'étude des processus morphodynamiques à l'échelle des bassins versants, des fonds de vallées et des lits fluviaux, ce travail vise à répondre à la problématique générale suivante : Dans quelle mesure le fonctionnement actuel et sub-actuel des rivières vosgiennes est-il contrôlé par les héritages (fluvio-)glaciaires et anthropiques ?

L'étude se décline en trois grands objectifs, qui correspondent aux chapitres 2, 3 et 4 du manuscrit. Cette partie vise à problématiser chacun des objectifs à travers un court état de l'art et à préciser les méthodes déployées ainsi que les échelles spatio-temporelles impliquées. Les trois objectifs, les méthodes et les échelles spatio-temporelles associées sont déclinés dans le Tableau 1.1. La localisation des méthodes employées est cartographiée dans la Figure 1.12.

Le premier objectif est d'évaluer l'état d'équilibre sédimentaire du massif des Vosges et de mesurer les taux de dénudation des bassins versants vosgiens (Tableau 1.1; Fig. 1.12A). Le deuxième objectif est de reconstituer l'évolution morpho-sédimentaire postérieure au dernier maximum glaciaire (DMG) d'un système fluvial fortement impacté par les glaciations quaternaires (Tableau 1.1; Fig. 1.12B). Le troisième objectif est de reconstituer la trajectoire géomorphologique historique des lits fluviaux et de leurs fonds de vallées et d'évaluer la part respective des facteurs de contrôle, notamment anthropiques (Tableau 1.1; Fig. 1.12B).

Enfin, le Chapitre 5 propose une synthèse des nouvelles connaissances acquises grâce à ce travail et des perspectives spécifiques à chacun des trois grands objectifs (Chapitres 2, 3 et 4). De

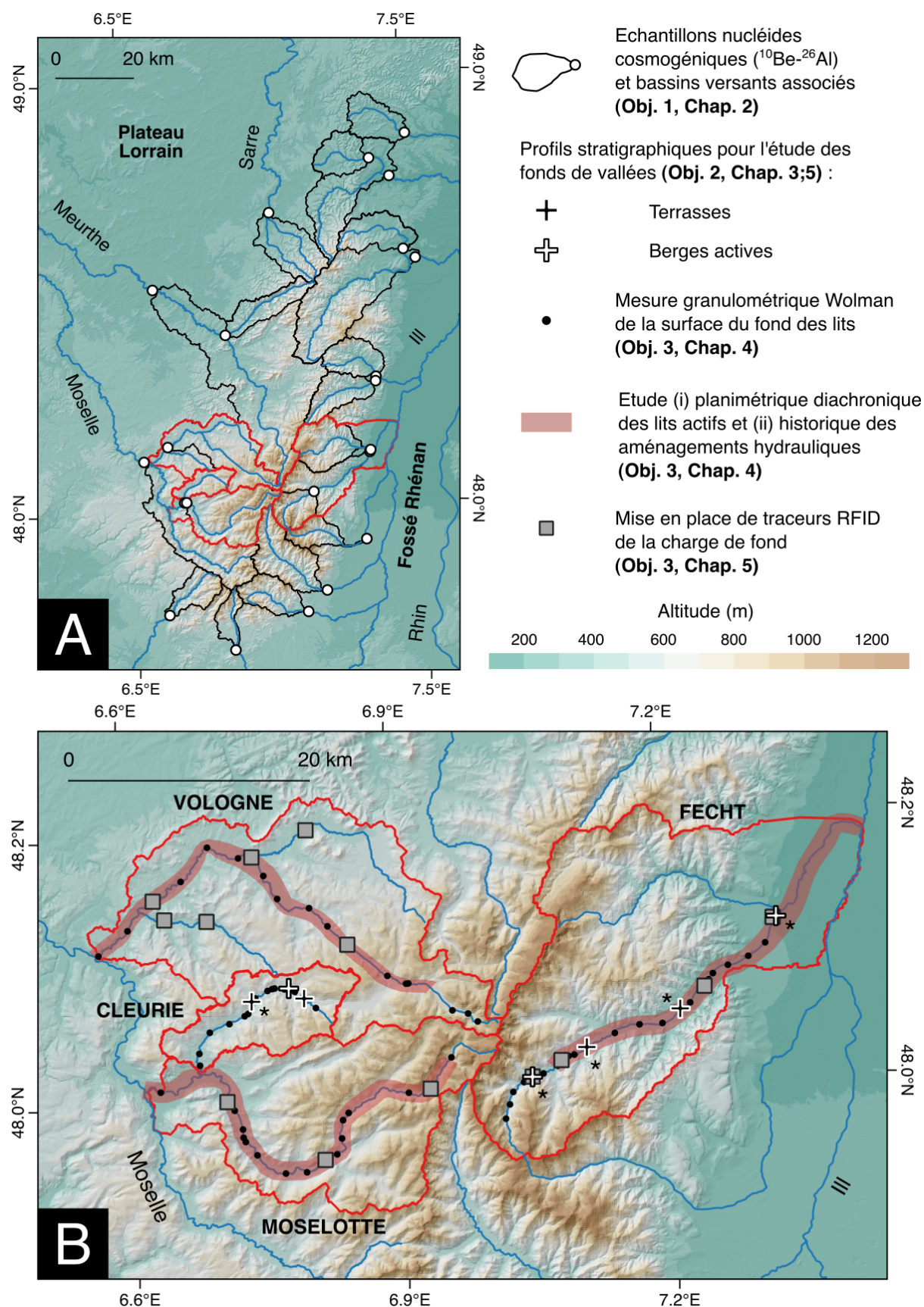


FIGURE 1.12: Localisation et détail des méthodes employées dans le cadre de cette étude, à l'échelle (A) du massif des Vosges et (B) du transect Est-Ouest composé des bassins versants de la Fecht, Moselotte, Vologne et Cleurie. Les profils stratigraphiques identifiés par une astérisque (*) sont présentés dans le Chapitre 5.

TABLE 1.1: Structure de la thèse, méthodes et échelles spatiales et temporelles impliquées.

	Grands objectifs	Méthodes	Échelle spatiale	Période visée
1	Évaluer l'état d'équilibre cosmogénique et mesurer les taux de dénudation des bassins versants vosgiens.	Mesure de la concentration en ^{10}Be et ^{26}Al dans des échantillons de sable, prélevés dans les lits fluviaux, au débouché du massif.	Entièrement du massif des Vosges	Holocène
2	Reconstituer l'évolution post-DMG d'un système fluvial fortement impacté par les glaciations.	Approche stratigraphique et morphométrique multi-méthodes (radiocarbone, stratigraphie, granulométrie, profils en long).	Bassin versant (Cleurie)	Holocène
3	Quantifier l'impact de la pression humaine sur la trajectoire géomorphologique des lits fluviaux.	Étude de l'évolution de l'occupation du sol, étude planimétrique diachronique du lit actif, spatialisation et datation des structures hydrauliques.	Transect Est-Ouest (Fecht, Moselotte, Vologne)	Moderne, Contemporaine

plus, dans la perspective d'un diagnostic géomorphologique fonctionnel pour une meilleure gestion des rivières vosgiennes, des premières pistes de réflexions voire de recommandations seront présentées. Ce chapitre présentera également des résultats préliminaires supplémentaires ayant été acquis dans le cadre ce travail. Ils concernent d'une part l'étude de quatre profils stratigraphiques de terrasses et de berges dans la vallée de la Fecht, et d'un profil d'une probable terrasse de kame dans la vallée de la Cleurie (Fig. 1.12B). D'autre part, nous présenterons quelques résultats préliminaires concernant le suivi du transport de la charge de fond *via* la mise en place de traceurs à radio-identification (Radio Frequency IDentification; RFID) dans les lits de la Fecht, la Moselotte, la Vologne et deux de ses affluents (Fig. 1.12B).

1.3.1 Dénudation et équilibre cosmogénique du massif des Vosges

La dénudation correspond au processus dominant dans la zone 1 du système fluvial (Fig. 1.1; Schumm, 1977). Elle regroupe l'altération chimique de la surface de la roche et l'érosion physique de cette dernière, c'est à dire le détachement et la mise en mouvement des particules (Dunai, 2010). La mesure des concentrations en nucléides cosmogéniques dans les sédiments quartzeux prélevés dans le lit d'un cours d'eau est désormais une méthode bien établie pour calculer le taux de dénudation moyen de son bassin versant (von Blanckenburg, 2005; Dunai, 2010), sur des échelles de temps allant de plusieurs milliers à plusieurs centaines de milliers d'années (Godard et al., 2012b; Rixhon, 2022). La méthode repose sur la relation inverse entre la concentration en nucléides cosmogéniques et le taux de dénudation. Plus la concentration

en nucléides cosmogéniques est élevée, plus longtemps le sédiment a été exposé en (sub-)surface aux rayonnements cosmiques, moins la dénudation est forte (Mariotti, 2020). Depuis le développement de la méthode au milieu des années 90 (Brown et al., 1995; Granger et al., 1996; Bierman and Steig, 1996), cette approche a été appliquée à des bassins versants couvrant pratiquement tout les contextes tectoniques et climatiques à travers le monde, y compris des secteurs actuellement (Grin et al., 2018) et anciennement (Buechi et al., 2014) englacés. En plus de pouvoir quantifier la dénudation pluri-millénaire d'un bassin versant, ou sa variabilité spatiale au sein d'un massif entier, ces études cherchent généralement à évaluer l'influence respective de certains facteurs de contrôle potentiels de la dénudation, comme la topographie (Peifer et al., 2021), la tectonique (Olivetti et al., 2016), le climat (Carretier et al., 2013) ou la lithologie (Cyr et al., 2014).

A travers les centaines d'études réalisées, deux constats importants peuvent être faits. D'abord, la capacité à évaluer l'influence de l'un ou plusieurs facteurs de contrôle sur la dénudation semble dépendante de la présence d'une forte variabilité en termes de gradients morphométriques (Hippe et al., 2012), tectoniques (Godard et al., 2014), climatiques (Moon et al., 2011; Ferrier et al., 2013) ou lithologiques (Kober et al., 2015). Ensuite, pour la grande majorité des quelques milliers d'échantillons collectés dans ces études, seule la concentration d'un nucléide (^{10}Be) a été mesurée, ce qui implique deux hypothèses fortes pour en déduire des taux de dénudation. D'une part, l'hypothèse d'état d'équilibre cosmogénique du bassin versant, qui revient à considérer que le flux entrant de nucléides cosmogéniques dans le système fluvial (*via* la production, zone 1; Fig. 1.3), est égal au flux sortant (*via* la dénudation, le transport et l'export des sédiments; zone 2-3; Fig. 1.3). D'autre part, la durée du transfert des sédiments jusqu'à l'exutoire (le temps de transit) doit être courte par rapport au temps pendant lequel la dénudation se produit (Dunai, 2010). Or, l'évaluation de l'état d'équilibre et du temps de transit est possible uniquement par la mesure conjointe de la concentration de deux nucléides aux demi-vies différentes (typiquement ^{26}Al et ^{10}Be). Cette approche a été largement utilisée pour mesurer l'efficacité de l'érosion glaciaire sur la roche mère dans des régions anciennement englacées (Corbett et al., 2013; Strunk et al., 2017; Knudsen and Egholm, 2018; Andersen et al., 2020). Les faibles ratios $^{26}\text{Al}/^{10}\text{Be}$ mesurés dans de nombreux échantillons de roche mère indiquent clairement que l'hypothèse d'état d'équilibre n'est pas vérifiée, en raison d'un enfouissement répété et étendu sous d'épaisses masses de glace. Pourtant, l'approche à deux nucléides (^{26}Al - ^{10}Be) n'a été appliquée que dans quelques études pour les sédiments de cours d'eau modernes, bien que Wittmann et al. (2020) aient récemment mis en évidence des ratios $^{26}\text{Al}/^{10}\text{Be}$ significativement appauvris dans 35 % des plus de 50 grands fleuves qu'ils ont étudiés à travers le monde. En Europe Occidentale, d'après la base de données OCTOPUS (consultée le 5 avril 2024; Codilean et al., 2022), la mesure conjointe des concentrations en ^{26}Al - ^{10}Be dans des sédiments de rivière n'a été effectuée que pour 1,8 % (13/734) des échantillons prélevés (Fig. 1.13). Ils correspondent aux mesures effectuées par

Wittmann et al. (2016) dans le bassin du Pô (Italie). Leurs résultats ont permis de valider l'hypothèse d'état d'équilibre dans les bassins étudiés. Concernant les 98,2 % d'échantillons restants, en revanche, l'hypothèse d'état d'équilibre n'a pas été testée. De plus, près de deux tiers d'entre eux ont été prélevés dans des cours d'eau dont les bassins versants sont ou ont été au moins partiellement englacés. C'est le cas notamment dans les Alpes et en Forêt-Noire (Fig. 1.13). Or, il a été montré qu'en contexte glaciaire, l'hypothèse d'état d'équilibre est particulièrement critique, que ce soit dans des bassins anciennement (Glotzbach et al., 2013) ou présentement (Delunel et al., 2014) englacés.

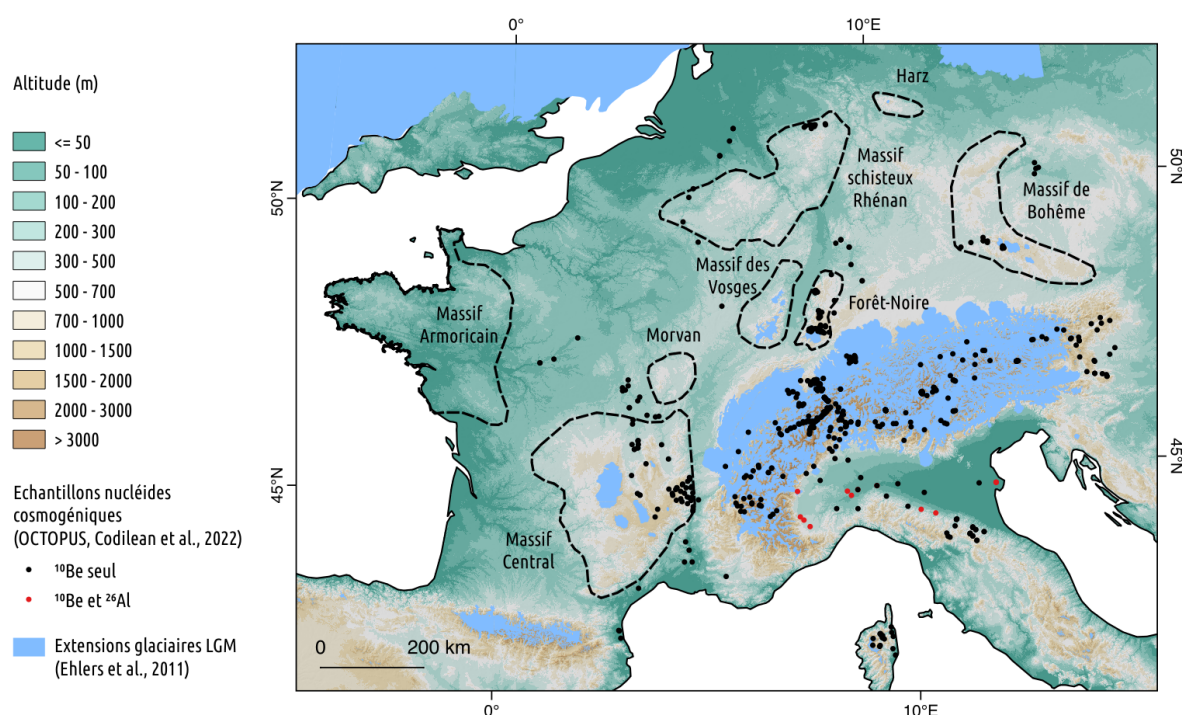


FIGURE 1.13: Localisation des massifs hercyniens en Europe Occidentale et compilation des échantillons de sédiments de rivières prélevés pour le calcul des taux de dénudations à partir des concentrations en nucléides cosmogéniques (^{10}Be seul ou combiné avec ^{26}Al), d'après la base de données OCTOPUS (Codilean et al., 2022, consultée le 5 avril 2024).

Ainsi, la problématique suivante peut être posée : A l'échelle d'un massif anciennement et partiellement englacé, est-il possible d'établir une relation entre le déséquilibre cosmogénique des bassins versants et leur degré d'englacement passé ? En tirant parti des caractéristiques contrastées du massif des Vosges (Fig. 1.9), l'étude du Chapitre 2 vise à (i) démontrer la structuration spatiale du massif en termes de potentiels facteurs de contrôle de la dénudation, (ii) évaluer le (dés)équilibre cosmogénique des bassins versants, (iii) explorer l'impact de la couverture glaciaire et du stockage de sédiments d'origine glaciaire sur les rapports de concentration cosmogéniques. Premièrement, des indices morphométriques et climatiques ont été calculés et des groupes de bassins versants lithologiquement uniformes ont été statistiquement déterminés. Deuxièmement, les sédiments du lit des cours d'eau drainant l'ensemble

des bassins versants du massif des Vosges ont été échantillonnés pour mesurer leur concentrations en ^{26}Al et ^{10}Be (Fig. 1.12A). Troisièmement, une comparaison systématique a été effectuée entre les rapports $^{26}\text{Al}/^{10}\text{Be}$ et la couverture glaciaire ancienne ainsi que la distribution spatiale des sédiments (fluvio-)glaciaires.

1.3.2 Ajustements d'un système fluvial fortement impacté par les glaciations : le bassin versant de la Cleurie

Les glaciations quaternaires ont largement contribué au façonnement ancien des paysages fluviaux actuels. La dynamique morpho-sédimentaire et l'état d'équilibre des systèmes fluviaux qui évoluent dans ces vallées anciennement englacées est ainsi conditionnée par les formes et dépôts glaciaires, qui en constituent les héritages (Cordier et al., 2017; Mason and Polvi, 2023). Jusqu'à présent, les études sur l'évolution post-glaciaire des paysages fluviaux dans les zones fortement marquées par les glaciations se sont principalement concentrées sur les chaînes de haute montagne (par exemple les Alpes européennes : Serra et al., 2022; Piermattei et al., 2023). Dans la grande majorité des cas, les processus fluviaux se voient attribuer un rôle central dans le façonnement ultérieur du paysage (Gomez and Livingston, 2012), notamment dans les contextes où des paléolacs se sont formés (Tielidze et al., 2021). En revanche, l'évolution post-glaciaire des systèmes fluviaux des chaînes de basse ou moyenne montagne anciennement englacées est relativement peu étudiée (Rixhon, 2021).

Le bassin versant de la Cleurie (environ 80 km²), situé au cœur de la zone autrefois englacée du massif des Vosges (Fig. 1.11), s'y distingue par plusieurs caractéristiques géomorphologiques singulières. Tout d'abord, en raison de la probable convergence de trois langues glaciaires (Flageollet, 1988, 2002), il présente la plus forte concentration de dépôts glaciaires et fluvio-glaciaires parmi tous les bassins versants autrefois englacés, occupant plus de 40 % de sa surface (Fig. 1.11; Jautzy et al., 2024b). Ensuite, ces dépôts sont particulièrement variés et incluent des dépôts morainiques, deltaïques et lacustres, dont l'origine serait liée à la présence d'un ou plusieurs paléolac(s) (Flageollet, 1988). Bien qu'un certain nombre de travaux précurseurs aient déjà exploré ce cadre géomorphologique singulier (Hogard, 1840; Seret, 1967; Salome, 1968; Flageollet et al., 1971; Flageollet, 1988, 2002, Fig. 1.14), aucun d'entre eux n'a étudié l'évolution fluviale post-glaciaire de la vallée de la Cleurie, ni n'a produit de datations numériques concernant les principales étapes de la (dé)glaciation de cette vallée singulière. Par ailleurs, les interprétations de ces auteurs concernant les processus responsables de la formation des principales formes glaciaires rencontrées dans la vallée ont souvent été divergentes. En particulier, Salome (1968), Seret (1967) et Flageollet (1988) ont respectivement interprété les principales formes glaciaires comme étant la conséquence d'une seule (Fig. 1.14B), de deux (Fig. 1.14C) et de trois (Fig. 1.14C) langues glaciaires.

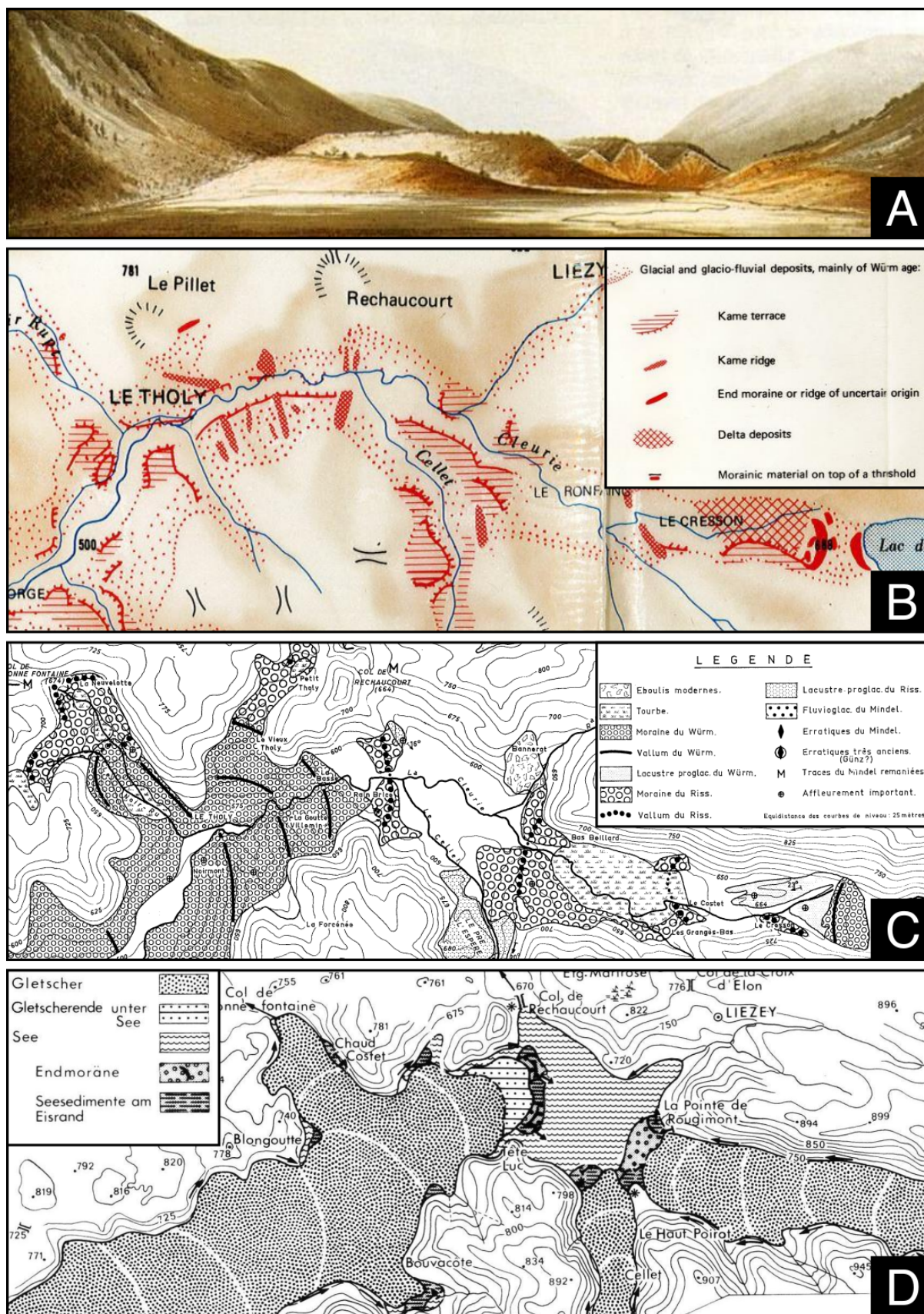


FIGURE 1.14: Extraits des premiers travaux géomorphologiques dans la vallée de la Cleurie. (A) Aquarelle représentant le complexe de moraines frontales incisées par le cours d'eau de la Cleurie, au niveau du Tholy (Hogard, 1840). (B) Carte géomorphologique des dépôts de versants et fonds de vallées, principalement attribués à des dépôts de kame. Une seule langue glaciaire est supposée (Salome, 1968). (C) Carte géomorphologique des dépôts de versants et fonds de vallées, principalement attribués à des dépôts morainiques. Deux langues glaciaires convergentes sont supposées (Seret, 1967). (D) Carte représentant l'extension des trois langues glaciaires supposées au dernier maximum glaciaire (Flagecollet, 1988).

A ce titre, le bassin versant de la Cleurie représente selon nous un cas d'étude exceptionnel à l'échelle du massif des Vosges pour caractériser et évaluer l'impact des héritages glaciaires sur les ajustements fluviaux post-DMG d'un cours d'eau de moyenne montagne à énergie modérée. La reconstitution de l'évolution post-glaciaire du système fluvial de la Cleurie doit permettre d'évaluer la capacité d'un cours d'eau vosgien à se réajuster à la suite d'une perturbation climatique majeure, à savoir le DMG. Cet aspect est d'autant plus pertinent dans le cadre de mise en place de stratégies de gestion voire de restauration des cours d'eau vosgiens. En effet, il est de plus en plus reconnu que les héritages glaciaires sont un facteur de contrôle majeur de la dynamique morpho-sédimentaire des cours d'eau, dont la prise en compte est essentielle (Mason and Polvi, 2023).

L'étude du Chapitre 3 vise donc à (i) évaluer l'état d'équilibre du système fluvial de la Cleurie et à (ii) reconstruire l'évolution fluviale post-DMG de la vallée. Premièrement, sur base des travaux géomorphologiques précurseurs entrepris entre le milieu du XIX^{ème} siècle (Hogard, 1840, 1842, Fig. 1.14A) et la fin du XX^{ème} siècle (Flageollet et al., 1971; Flageollet, 1988, Fig. 1.14D), de prospections de terrain et d'un modèle numérique de terrain (MNT) à haute résolution (1 m), une cartographie géomorphologique détaillée du fond de vallée a été réalisée. Deuxièmement, le degré d'ajustement du réseau hydrographique a été évalué à partir de l'analyse (i) des profils en long du chenal principal et des principaux affluents, et (ii) de l'évolution longitudinale de la granulométrie de surface du fond du lit de la Cleurie, dont 1400 particules ont été mesurées (Wolman, 1954). Enfin, deux profils sédimentaires clés en fond de vallée ont été étudiés (Fig. 1.12B), à partir notamment de mesures granulométriques et de datations par radiocarbone de matière organique.

1.3.3 Quantification et impact de la pression humaine sur la trajectoire géomorphologique de la Fecht, de la Moselotte et de la Vologne

Au cours des derniers siècles, la morphodynamique des systèmes fluviaux dans la plupart des régions du monde a été considérablement impactée par la pression humaine (Wohl, 2020; Surian, 2022; Brenna et al., 2024), qui est désormais reconnue comme un facteur majeur des changements fluviaux modernes (Fig. 1.3; Liébault and Piégay, 2002; Salit et al., 2015; Verstraeten et al., 2017; Cooper et al., 2018). Elle correspond à des interventions humaines directes et/ou indirectes à l'échelle du bassin versant, du fond de vallée et du lit du cours d'eau, telles que les changements d'usages du sol (Liébault and Piégay, 2002; Delile et al., 2016; Jautzy et al., 2021), la régulation/modification des débits (Winterbottom, 2000), l'extraction de granulats (Comiti et al., 2011; Ziliani and Surian, 2012), ainsi que la chenalisation, l'endiguement et la construction de barrages et de seuils (Arnaud et al., 2019; Mandarino et al., 2019). Avec seulement un tiers des rivières européennes répondant aux critères de 'bon état écologique' selon la DCE, les systèmes fluviaux européens sont probablement parmi les plus dégradés au

monde (Belletti et al., 2020), ce qui est lié à la longue histoire d’occupation et d’exploitation des hydrosystèmes fluviaux européens. De nombreuses études ont mis en évidence la relation entre l’anthropisation des versants, des fonds de vallée et des chenaux et les changements fluviaux observés, notamment au cours des deux derniers siècles (Ziliani and Surian, 2012; Salit et al., 2015; Arnaud et al., 2019). Étant donné que la dynamique fluviale, y compris les formes et les processus associés, conditionnent les habitats, la biodiversité et les services écosystémiques, toute altération des processus morpho-sédimentaires ou des morphologies du chenal conduit inévitablement à une perte d’habitats, une réduction de la biodiversité et une diminution des services écosystémiques (Lespez et al., 2015; Brown et al., 2018). Ces dernières décennies, la reconnaissance croissante de l’impact de la pression humaine sur la dégradation des systèmes fluviaux européens a conduit à de nombreux programmes de restaurations (García et al., 2021).

À cet égard, l’objectif principal de la DCE est le retour des masses d’eau de surface, dont les cours d’eau, à un ‘bon état écologique’ (Schmitt et al., 2007; Dufour and Piégay, 2009; Gob et al., 2014; Maaß et al., 2021), lui même lié à la notion d’état de ‘référence’, voire d’état ‘naturel’. Or, la pertinence de la notion d’état de référence est largement critiquée compte tenu de l’ancienneté et du poids des actions humaines, directes et/ou indirectes, sur les cours d’eau et leurs bassins versants (Bravard, 2003; Dufour and Piégay, 2009; Brown et al., 2018). Lorsque l’on considère la restauration d’un cours d’eau, il est donc essentiel de prendre en compte le fonctionnement actuel du cours d’eau, qui conditionne les possibilités de restauration. Un préalable essentiel à la caractérisation du fonctionnement actuel du cours d’eau est donc d’évaluer la part respective des différents facteurs de contrôle qui ont conduit à ce fonctionnement passé, et/ou qui le maintiennent dans cet état (Liébault, 2003).

Cependant, en raison de la nature non linéaire des interactions entre les facteurs anthropiques et naturels, interpréter les relations de cause à effet dans les systèmes fluviaux affectés par de multiples interventions humaines, qu’il s’agit de déconvoluer, est une tâche délicate (Downs et al., 2013; Brenna et al., 2024). De plus, la récente revue de littérature menée par Downs and Piégay (2019) sur les études des facteurs de contrôle humains sur les changements fluviaux a montré que les tentatives de démonstration des relations de cause à effet proposées par la plupart des auteurs sont souvent insuffisamment robustes ou manquent d’arguments scientifiques convaincants et de quantification précise. Premièrement, les forçages humains, qu’ils soient directs ou indirects, sont rarement précisément spatialisés. Deuxièmement, l’impact cumulatif résultant de multiples facteurs anthropiques potentiels est généralement négligé. Troisièmement, la chronologie des événements est souvent partielle ou plutôt vague (Verstraeten et al., 2017).

Dans le Chapitre 4, nous nous focalisons sur trois bassins versants adjacents de la partie sud du massif des Vosges, les bassins versants de la Fecht, de la Moselotte et de la Vologne (Fig.

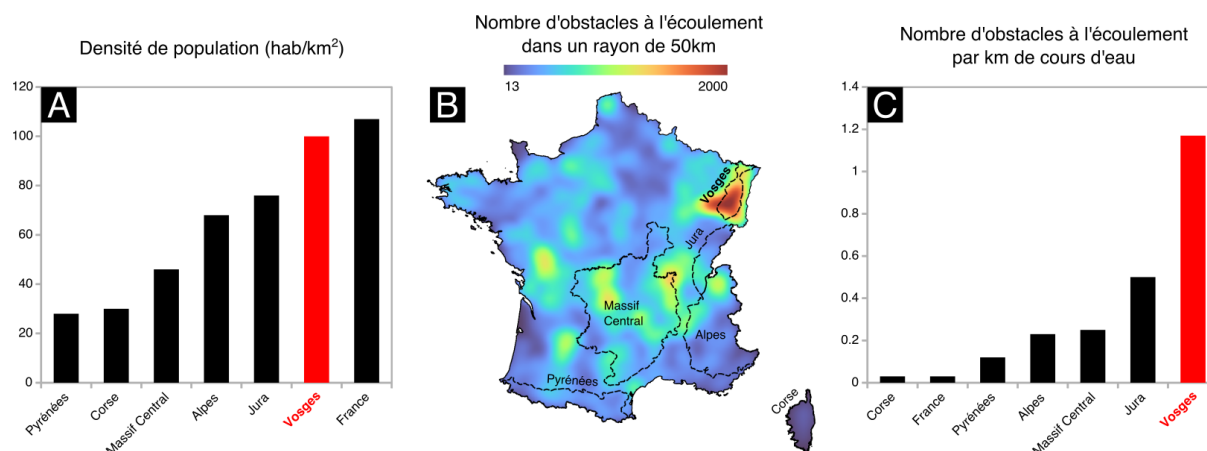


FIGURE 1.15: (A) Densité de population dans les principaux massifs montagneux français, d'après la base de données Filosofi2017 (INSEE; <https://www.insee.fr/fr/statistiques/6215138?sommaire=6215217>). (B) Carte de chaleur représentant le nombre d'obstacles à l'écoulement dans un rayon de 50 km, d'après le Référentiel national des Obstacles à l'Écoulement (ROE, 2019). (C) Nombre d'obstacles à l'écoulement par kilomètre de cours d'eau dans les principaux massifs montagneux français. Les massifs sont délimités d'après l'Observatoire des Territoires (<https://www.observatoire-des-territoires.gouv.fr/kiosque/zonage-les-perimetres-de-massifs>).

1.8). Le massif des Vosges a été significativement impacté pendant l'ère de l'industrialisation pour produire l'énergie hydraulique nécessaire au fonctionnement de diverses industries (principalement industries textiles, papeteries et scieries) et pour irriguer les terres cultivées (Parmentier, 2019). En conséquence, de tous les massifs français, le massif des Vosges et son réseau hydrographique affichent respectivement la plus forte densité de population (Fig. 1.15A) et le plus grand nombre moyen d'obstacles à l'écoulement par kilomètre de cours d'eau (Fig. 1.15C).

Le principal objectif de l'étude du Chapitre 4 est d'évaluer dans quelle mesure l'anthropisation des bassins versants, des fonds de vallées et des chenaux de la Fecht, de la Moselotte et de la Vologne, a affecté le style fluvial et la dynamique des cours d'eau durant les deux derniers siècles. Parallèlement, nous testons l'intérêt de l'usage d'archives documentaires pour attribuer des périodes de construction aux seuils présents dans les lits des rivières. Nous tirons parti des archives documentaires produites par l'administration des Ponts et Chaussées, qui contiennent toutes les requêtes de construction ou de régulation de structures hydrauliques le long des rivières françaises, du début du XIX^{ème} siècle à la fin du XX^{ème} siècle (Payraastre, 2005). Bien que rarement utilisées, quelques études ont déjà démontré l'intérêt de leur usage dans la reconstruction géohistorique fluviale sur les deux derniers siècles (Jacob-Rousseau, 2015; Mesmin et al., 2024).

Premièrement, sur la base du Référentiel national des Obstacles à l'Écoulement (ROE, 2019), d'un MNT à haute résolution (1 m) et de la prospection approfondie des archives documentaires de l'administration des Ponts et Chaussées, une base de données originale a été construite, recensant l'intégralité des seuils, digues et ponts présents dans le lit et le fond de vallée

des trois cours d'eau. Deuxièmement, l'évolution planimétrique diachronique des lits actifs a été reconstruite à partir de la carte d'état-major (1832) et de deux orthophotographies (1951, 2018). Troisièmement, l'influence potentielle des facteurs de contrôle géomorphologiques a été évaluée en mesurant l'évolution longitudinale de la largeur du fond de vallée, de la présence de dépôts (fluvio-)glaciaires et de la pente du talweg. Les patrons longitudinaux de la granulométrie de surface du fond des lits actifs (~5000 particules) ont également été mis en évidence. Enfin, à l'échelle des bassins versants, l'évolution de l'occupation du sol entre le début du XIX^{ème} et le début du XXI^{ème} siècle a été mesurée à partir de la digitalisation de la carte d'état-major et des données actuelles (OCSGE, Région Grand Est).

Chapitre 2

Cosmogenic (un-)steadiness revealed by paired-nuclide catchment-wide denudation rates in the formerly half-glaciated Vosges Mountains (NE France)

This article has been published in the journal **Earth and Planetary Science Letters** (Jautzy et al., 2024b). Results were presented as a poster at **MGM2021** (Munich, Germany), and as an oral at **EGU2021** (online), **Goldschmidt2021** (online), **ICG2022** (Coimbra, Portugal) and **Quaternaire13, 2022** (Strasbourg, France).

Article abstract Although catchment-wide denudation rates inferred from *in situ* cosmogenic nuclide concentrations measured in stream sediments has represented a ground-breaking progress in geomorphology over the last three decades, most of these studies rely on ^{10}Be concentrations only. It seems that this current and routine one-nuclide approach to infer catchment-wide denudation rates has somehow overshadowed two key assumptions that are cosmogenic steady-state and short sediment transit time at the catchment scale. Although a paired-nuclide approach allow testing these assumptions, it is rarely performed on stream sediments and this can become highly problematic in slow-eroding, formerly glaciated contexts. In this study, we thus measure both ^{10}Be and ^{26}Al in stream sediments pertaining to twenty-one rivers draining an entire low mountain range: the Vosges Massif (NE France). The latter exhibits a sharp gradient between its southern and northern part in terms of lithology,

morphometry and climate. Moreover, if its northern part remained void of glacial cover during Quaternary cold stages, its southern part was significantly and repeatedly glaciated. We aim to assess the factors that control the denudation of the Vosges Mountains and to quantitatively explore the impact of both repeated glacial cover and storage of glacially derived sediments on $^{26}\text{Al}/^{10}\text{Be}$ ratios, hence cosmogenic (un-)steadiness in modern river samples. Our results first show that elevation, slope, channel steepness and precipitation are primarily organised along a N-S increasing trend. ^{10}Be - and ^{26}Al -derived catchment-wide denudation rates accordingly range from 34 ± 1 to 66 ± 2 , and 41 ± 3 to 73 ± 7 mm/ka, respectively, in thirteen investigated catchments that are in cosmogenic equilibrium. Lithological contrasts may control the pattern of denudation with a higher erodibility of the sandstone-dominated catchment to the north compared to the crystalline-dominated catchments to the south. Our results also show that catchments in strong cosmogenic disequilibrium ($^{26}\text{Al}/^{10}\text{Be}$ ratios from 1.4 to 5.2) spatially cluster in the SW part of the Vosges Mountains that was the most intensively glaciated during Quaternary cold stages. If this precludes any conclusion about controlling factors at the whole massif scale, this study is the first to quantify the impact of past glaciations on cosmogenic (un-)steadiness measured in stream sediments. A statistically significant relationship between the degree of depletion of the $^{26}\text{Al}/^{10}\text{Be}$ ratios and the spatial pattern of glaciation is found: the larger the former glacial cover in each catchment, the lower the $^{26}\text{Al}/^{10}\text{Be}$ ratio. Equally important is the significant correlation reported between the degree of depletion of the $^{26}\text{Al}/^{10}\text{Be}$ ratios and the proportion of glacial and fluvio-glacial deposits within each catchment. These two relationships underline the link between cosmogenic unsteadiness in the stream cosmogenic signal and long-lasting and repetitive ice shielding, and complex sediment routing systems in glacial environments, respectively. We thus argue to systematically measure ^{26}Al in complement to ^{10}Be and to test the steady-state assumption when it comes to infer catchment-wide denudation rates from modern stream sediments, especially in slow eroding, formerly glaciated landscapes.

2.1 Introduction

Assessment of catchment-wide denudation (CWD) rates inferred from *in situ* cosmogenic nuclide concentrations measured in modern stream sediments has represented a groundbreaking progress in geomorphology over the last three decades (von Blanckenburg, 2005; Dunai, 2010; Granger and Schaller, 2014). This approach has been applied to fluvial catchments from practically every tectonic and climatic setting worldwide, including currently (Grin et al., 2018) and formerly (Buechi et al., 2014) glaciated terrains. The key research aim of those studies consists in disentangling the complex interplays between topography (Peifer et al., 2021), tectonics (Olivetti et al., 2016), climate (Carretier et al., 2013), and lithology (Cyr et al., 2014) which primarily control denudation. The thousands of river-sediment samples

collected in the majority of case studies, however, were processed to measure *in situ* ^{10}Be concentrations only (Portenga and Bierman, 2011). Although this single-nuclide approach relies on several key assumptions, it seems that the current ^{10}Be 's routine approach used to infer CWD rates has somehow overshadowed two of them. First, little attention has been recently paid to cosmogenic steady-state at the catchment-scale (von Blanckenburg, 2005). It implies that the catchment's inbound flux *via* ^{10}Be production is counterbalanced by the outbound flux by denudation and decay at the catchment's outlet where the sampling takes place. Second, the timespan over which sediments are transported within the catchment to the outlet, i.e. sediment transit time, must be short compared to the timespan over which denudation occurs (Dunai, 2010). In settings where these conditions might not be met, ^{10}Be radioactive decay can become sizeable so that special care must be taken in settings where long-term sediment storage and remobilisation is evidenced. Neither the assumption of catchment's steady-state nor that of short sediment transit time can be checked by processing river-sand ^{10}Be concentrations only. As for steady-state, glacier cover, glacial erosion and/or glacially derived sediments may significantly impact the estimation of ^{10}Be -derived CWD rates in currently (Delunel et al., 2014) or formerly (Glotzbach et al., 2013) glaciated catchments. As for transit time, glacial and post-glacial environments have complex sediment routing systems (Porter et al., 2019) where reworking of ancient sediments impacts cosmogenic inventories as well (Wittmann and von Blanckenburg, 2009; Buechi et al., 2014). A few ^{10}Be -based studies developed some modelling approaches to partially cope with this twofold key issue (Godard et al., 2012a; Glotzbach et al., 2013).

Examination of steady-state, transit time and potential storage are potentially better constrained by measuring both ^{10}Be and ^{26}Al concentrations in the same quartz-bearing sample (Bhattacharjee et al., 2023). Since its original formulation in Lal's 1991 paper, the two-nuclide diagram distinguishes those samples that meet the steady-state assumption by plotting in the so-called steady-erosion island (Dunai, 2010) from those that underwent complex exposure histories and are characterised by depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios. This approach has been widely used in formerly glaciated regions of e.g. Greenland (Corbett et al., 2013; Strunk et al., 2017; Andersen et al., 2020) and Scandinavia (Andersen et al., 2018; Knudsen and Egholm, 2018) to successfully assess glacial erosion efficiency and denudation rates. Depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios in many glacial bedrock and erratic samples clearly violate the steady-state assumption due to repeated and extensive burial under thick ice masses with low erosive potential, i.e. cold-based glaciers. On the contrary, the paired-nuclide (^{26}Al - ^{10}Be) approach was only applied in few studies for modern stream sediments, although Wittmann et al. (2020) recently highlighted significantly depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios in 35 % of the >50 large rivers they studied all over the Earth's surface. Merely a handful of these studies dealt with glaciated catchments (Safran et al., 2005; Hippe et al., 2012; Grin et al., 2018; Wittmann et al., 2020; Zhang et al.,

2021), but cosmogenic disequilibrium is highlighted not only in currently glaciated catchments of Tibet (Zhang et al., 2021) but also in formerly and marginally glaciated catchments of the Bolivian Andes (Hippe et al., 2012). However, Zhang et al. (2021) could only conclude that sediment reworking is a potential issue without further exploration nor quantification on this process. This altogether demonstrates our current lack of understanding in how cosmogenic signals measured in modern stream sediments might be perturbed in (formerly) glaciated catchments.

Located in northeastern France, the Vosges Massif (or Vosges Mountains; VM) is a favourable setting to investigate this question. This S-N oriented low-elevation mountain range exhibits a sharp dissimilarity between its southern and northern part in terms of lithology, morphometry and climate. Topographic imprints from successive Quaternary glaciations in the VM landscape also follow this spatial pattern: whereas most of the crystalline southern VM were repeatedly covered and shaped by thick glaciers (Flageollet, 2002; Mercier, 2014), the northern half was void of ice cover (Darmois-Théobald, 1972; Heyman et al., 2013). Taking advantage of the VM setting and characteristics, this study aims at (i) demonstrating the spatial structuration of the VM in terms of first-order controlling factors for catchment-scale denudation, (ii) evaluating the cosmogenic (dis)equilibrium in the VM and (iii) quantitatively exploring the impact of both repeated glacial cover and storage of glacially derived sediments on paired cosmogenic ratios. First, morphometric and climatic indices are computed and groups of lithologically uniform catchments are statistically determined. Second, modern stream sediments from 21 river-catchments outlets draining the whole VM are sampled for *in situ* ^{10}Be and ^{26}Al concentration measurements. Third, a systematic comparison is undertaken between $^{26}\text{Al}/^{10}\text{Be}$ ratios and high-resolution data including previous glacial coverage, as well as the spatial distribution of glacial and fluvio-glacial sediments.

2.2 Study area

The VM form the western footwall block of the southern Upper Rhine Graben (north-eastern France; Fig. 2.1A). Its uplift is assigned to Miocene lithospheric folding resulting from NW-directed compressional stresses exerted in the Alpine foreland (Ziegler and Dèzes, 2007). Plio-Quaternary tectonic re-activation and uplift were assumed as well (Harmand and Cordier, 2012). The VM's lithology primarily is bipartite: Paleozoic crystalline rocks and Triassic sandstones compose the southern ($\sim 2/3$) and northern ($\sim 1/3$) parts, respectively (Fig. 2.1B). Strongly contrasting with the almost uniform Mesozoic sedimentary cover to the north, the Paleozoic basement reaches Hercynian orogenic units with various lithologies (Fig. 2.1B; Skrzypek (2011)). Plutonic rocks prevail (i.e. granites and (grano-)diorites mostly) but metamorphic (i.e. mostly gneiss) and volcanic (i.e. various lavas and tuffs) rocks also occur.

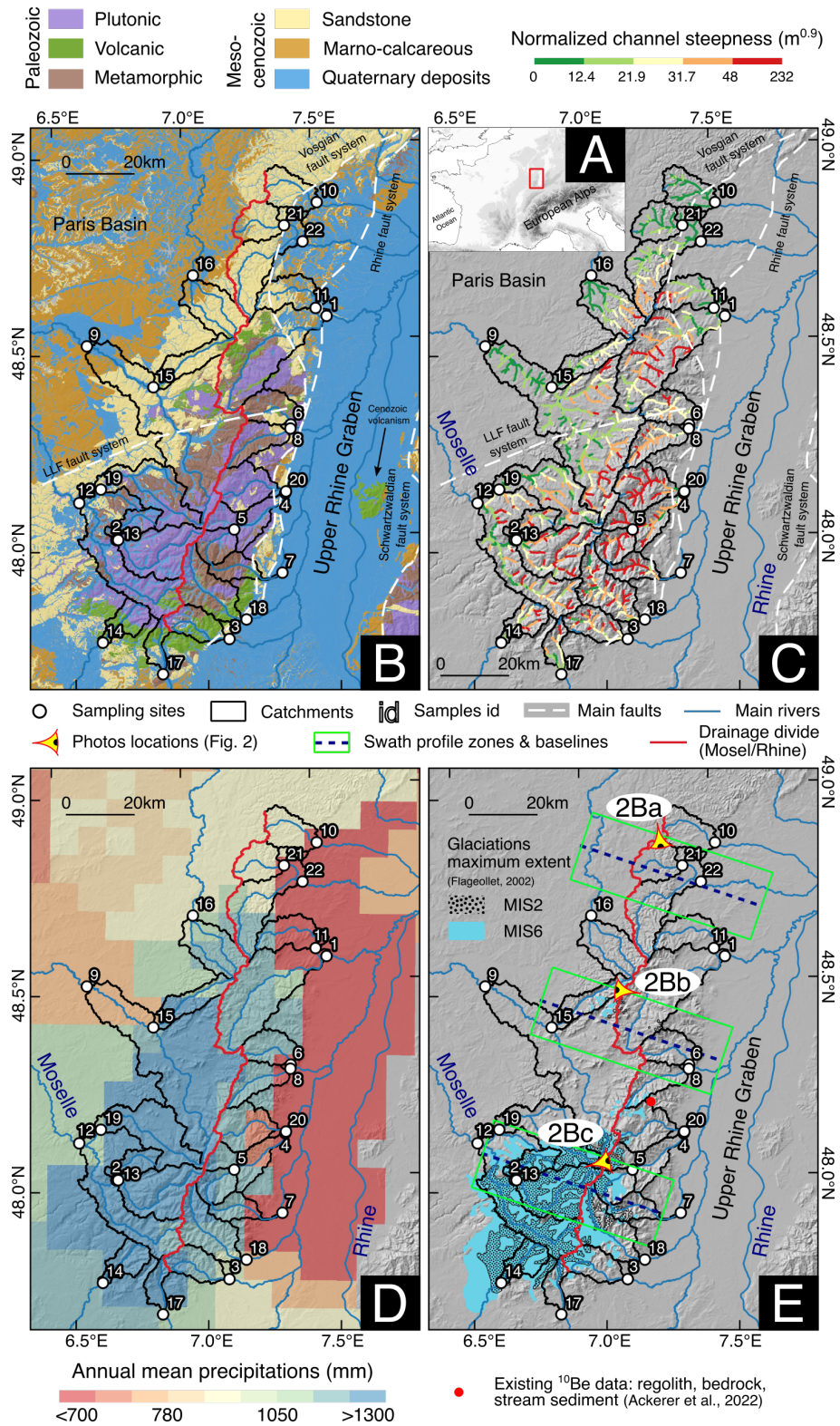


FIGURE 2.1: Main characteristics of the Vosges Mountains (VM) with sample locations for cosmogenic nuclide analysis and their associated catchments. (A) Location of the VM in Central Europe, forming the southwestern uplifted flank of the Upper Rhine Graben. (B) Simplified geological map (1/50.000) according to the French Geological Survey (BRGM). (C) Normalized channel steepness index of the hydrological network calculated in the sampled catchments using LSD Topotools software (Mudd et al., 2018) with a reference concavity of 0.45. (D) Annual mean precipitation from the national hydroclimatic database of Météo-France (Vidal et al., 2010) over the 1958–2009 period. (E) Glaciation maximum extents of the supposedly Marine Isotope Stage (MIS)2 and MIS6 cold periods (Flageollet, 2002). Green rectangles and yellow eyes match the swath profiles and the field pictures of Figure 2.2, respectively. The red point pinpoints already existing catchment-wide denudation rates in the VM (Ackerer et al., 2022).

The VM exhibit a well-marked N-S topographic gradient, from a low- to medium-elevation mountain range from north to south. Mean and peak elevations along three WNW-ESE oriented swath profiles performed in the VM's northern, central and southern parts amount to $\sim 255/525$, $470/1090$ and $710/1350$ m asl, respectively (Fig. 2.2A). A southward massif-scale increasing trend in normalised channel steepness index reflects this gradient as well (Fig. 2.1C). The VM is sharply bounded by the Vosgian normal fault system bordering the Upper Rhine Graben whereas its western side falls upon the slightly tilted Mesozoic formations from the Paris Basin (Figs. 2.1, 2.2A). Owing to the NNE/SSW- orientation of its main topographic divide, the VM typically form an almost perpendicular orogenic barrier to the predominant oceanic western atmospheric flow, thereby inducing a typical rain shadow effect (Fig. 2.1D; François and Humbert (2000)). Mean annual rainfall concomitantly increases with elevation along the main ridge as well, ranging from about 900 mm/a in the northern part to up to 1400 mm/a at elevations higher than 1350 m (Fig. 2.1D). Deciduous, resinous and mixed forests cover no less than $\sim 70\%$ of the VM's area (Rochel, 2004). They essentially occur on hillslopes and plateau-top positions, regardless the geographic location within the massif (except the bare, highest interfluvies in the southern part; Fig. 2.2Bc).

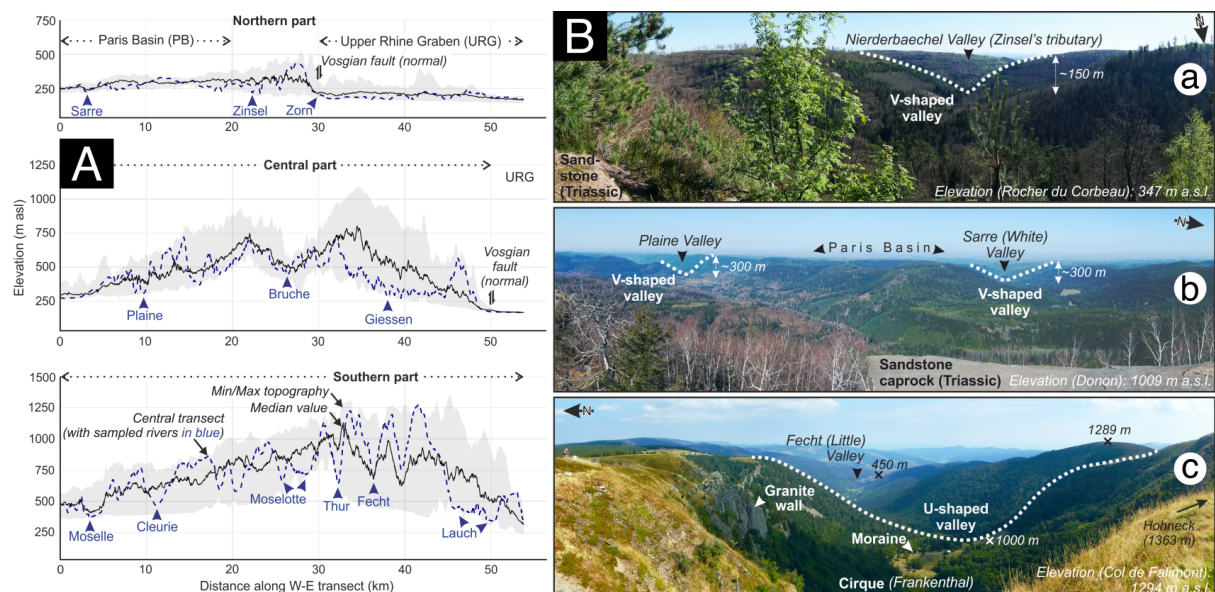


FIGURE 2.2: (A) North-South topographic gradient of the VM from swath profiles and (B) associated landscapes (locations in Fig. 2.1E).

The spatial pattern of Quaternary glaciations and their imprints on the landscape reflect this massif-scale dichotomy (Fig. 2.1E; Flageollet, 2002; Mercier and Jeser, 2004; Mercier, 2014). To the north, roughly two thirds of the VM exhibit fluvially overprinted landscapes (Fig. 2.2Ba-b). They completely lack evidence of former glacial activity and allow inferring an ice-free area in the northern third of the VM (Flageollet, 2002). Fluvial landscapes of the central part bear witness of local and sparse evidence of low-magnitude past glacial activity, i.e. small-dimensioned cirques and cirque moraines (Darmois-Théobald, 1972). Whilst mean fluvial

incision usually ranges from ~ 150 (Fig. 2.2Ba) to 300 m (Fig. 2.2Bb) in the northern and central third, respectively, this whole area most probably experienced periglacial conditions during Quaternary cold periods. By contrast, the VM's southern third (with mean elevations > 700 m) exhibit glacially overprinted landscapes. Numerous well-developed cirques along the main topographic divide, U-shaped valleys, glacial deposits (e.g. moraines; Fig. 2.2Bc), along with glacial and fluvio-glacial deposits blanketing the valley-floors (Supplementary Material, Fig. A.11) all indicate large-scale, past glacial activity. Despite the lack of numerical dating on glacial landforms, these landscapes bear witness of several Quaternary glaciations characterised by various ice extents (Fig. 2.1E). Flageollet (2002) recognised at least three to five glaciations. Spatial extensions of the supposedly Last Glacial Maximum (Marine Isotope Stage – MIS2) and the penultimate glaciation (MIS6) are judged reliable because they are based on abundant field evidence, contrasting with the poorly constrained pre-MIS6 glaciations (Supplementary Material, Fig. A.12). However, the study of sedimentary formations located at the VM's south(west)-ern margin lead Ménillet (2019) to consider the occurrence of Vosgian glaciations during the Early Pleistocene whose remains were erased during subsequent cold episodes. We thus suggest that the VM has repeatedly experienced significant glaciations over the whole Quaternary.

2.3 Data and methodology

2.3.1 Lithology, morphometry and climate

Massif-scale lithology was characterized using the 1:50,000 vectorized geological map provided by the French Geological Survey (BRGM; <https://infoterre.brgm.fr/formulaire/telechargement-cartes-geologiques-departementales-150-000-bd-charm-50>, last accessed 29th January 2021). We performed a query-attribute reclassification to obtain five main lithologies for bedrock units: plutonic, metamorphic, volcanic, and two classes in sedimentary rocks, i.e., sandstone (mostly Triassic sandstones) and marno-calcareous (at the VM's margins). Quaternary deposits were treated separately and likewise subdivided into alluvial, colluvial, and glacial/fluvio-glacial sediments. The respective proportions of each rock and deposit class were extracted within each sampled catchment (Fig. 2.3). Hierarchical clustering analysis was applied to our dataset of 21 catchments based on the aforementioned lithological classes. Further details on the method are given in Supplementary Material (A.1).

We used the 25 m-resolution digital elevation model (DEM) provided by the French National Institute of Geography (BD ALTI) to compute three commonly used catchment-averaged metrics: elevation, slope, and channel steepness index. For each DEM pixel, slope was computed as the steepest gradient between the eight neighboring pixels, using the function `terrain`

of the raster package in R (Hijmans, 2022). Normalized channel steepness index (k_{sn}) was computed using the LSD Topotools software (Mudd et al., 2018) with a reference concavity of 0.45 for inter-catchment comparison (Fig. 2.1C). We finally computed the mean annual precipitation over the VM (8-km resolution; Fig. 2.1D) from monthly-cumulated precipitation over the 1958-2009 period extracted from the national hydroclimatic database of Météo-France (Vidal et al., 2010). Catchment-averaged values for each topographic and climatic metric (i.e., elevation, slope, channel steepness, precipitation) were finally computed within each studied catchment (Table 2.1).

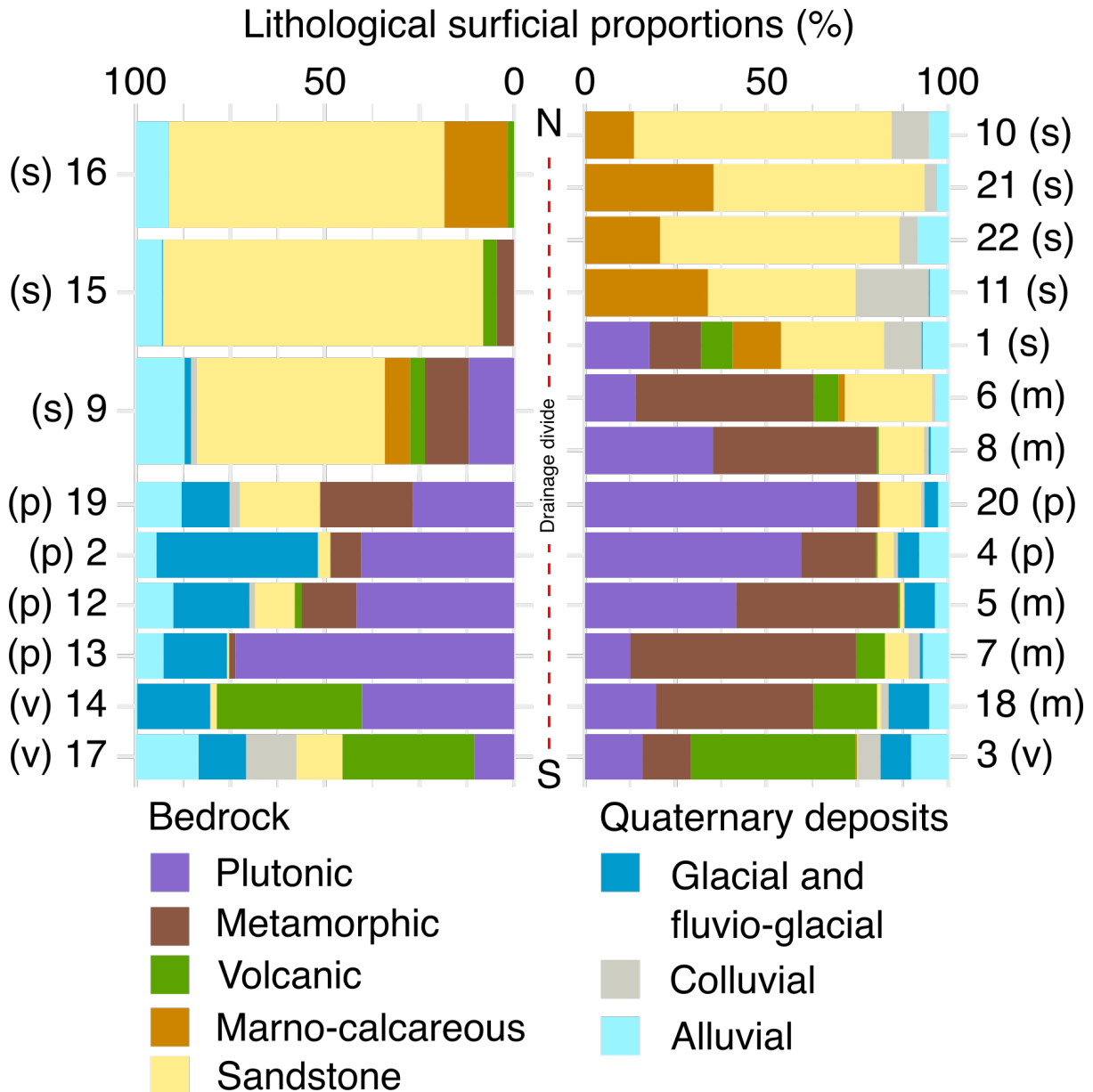


FIGURE 2.3: Lithological surficial proportions of each sampled catchments according to the 1/50.000 BRGM geological map. Catchments are classified according to their centroid's latitude. Letters in brackets corresponds to the results of the hierarchical clustering analysis performed on the bedrock lithologies and Quaternary deposits (see Section 2.3.1): sedimentary (s), metamorphic (m), plutonic (p), and volcanic (v).

2.3.2 Sampling strategy, AMS measurements and denudation rates

We sampled 22 stream sediments from 21 distinct rivers (two samples for the Fecht River; Table 2.1). Our sampling strategy covers all main rivers draining the VM along with some of their intra-massif tributaries (Fig. 2.1). Individual catchment sizes range from 62 to 1037 km² and the total surface considered in this study represents ~4,650 km² (Fig. 2.1). Sampling sites were selected at the outlet of the rivers' mountainous reach. Buechi et al. (2014)'s sampling strategy, primarily focusing on fluvially overprinted landscapes over glacially overprinted landscapes, was adapted with slight modifications (Figs. 1E, 2): the northern and southern parts of the sampled catchments overall match the fluvial landscapes and the glacial and fluvio-glacial landscapes, respectively.

River-sand samples were dried and sieved at the 250-1000 µm grain-size fraction. All ²⁶Al and ¹⁰Be chemical extractions and were performed at LN2C (National Laboratory of Cosmogenic Nuclides) located at the Centre Européen de Recherche et d'Enseignement des Géosciences de l'Environnement (CEREGE) laboratory in Aix-en-Provence (France) following standard procedures (e.g., Braucher et al. (2011); see Supplementary Material A.1 for further details). Accelerator Mass Spectrometry (AMS) measurements took place at the French national Facility, ASTER (Arnold et al., 2010). Measurement calibrations of the beryllium and aluminium measurements were done using the STD-11 standard (¹⁰Be/⁹Be ratio of $1.191 \times 10^{-11} \pm 1.08\%$; Braucher et al. (2015)) and the ASTER in-house standard SM-AL-11 (²⁶Al/²⁷Al ratio of $7.401 \pm 0.064 \times 10^{-12}$; Merchel and Bremser (2004)), respectively. The reported analytical uncertainties (1σ) include both the AMS counting statistics and internal error (0.5% for ¹⁰Be; Arnold et al. (2010)), chemical blank measurement and ²⁷Al measurements. ¹⁰Be/⁹Be blank ratio was 2.15×10^{-15} and ²⁶Al/²⁷Al blank ratio was less than 7.02×10^{-16} .

CWD rates were inferred from ¹⁰Be and ²⁶Al concentrations using the software CosmoCalc version 3.1 (<https://www.ucl.ac.uk/~ucfbpve/cosmocalc/>; Vermeesch, 2007), based on the time-independent Lal (1991)/Stone (2000) scaling model. The 22 catchments were first delineated using the 25-m resolution DEM and standard hydrological routing procedures. Latitude of each catchment's centroid as well as their average elevation were then calculated. For the sea-level-high-latitude (SLHL) reference production rates of ¹⁰Be and ²⁶Al, we used the default values implemented in CosmoCalc version 3.1 which are 4.3 and 30.95 at/g/a, respectively. Muonic productions was based on the scheme of Braucher et al. (2011). We assumed a standard bedrock density of 2.7 g/cm³ and did not apply any production correction for topographic shielding as reported by DiBiase (2018), nor for snow shielding.

2.4 Results

2.4.1 Lithological, morphometric and climatic spatial patterns

The spatial distribution of the different bedrock lithologies and Quaternary deposits reveals significant spatial variations over the VM and the 22 sampled catchments (Fig. 2.3). Some catchments are clearly dominated by a single main lithology, such as plutonic (e.g., #13, 20) and metamorphic (e.g., #7) bedrocks or sandstone (e.g., #10, 15). By contrast, other catchments are either constituted of two (e.g., #5, 14) or more (e.g., #1, 19) dominant lithologies. The hierarchical clustering analysis results in four catchments classes that are named after their main respective lithological composition: metamorphic (m), plutonic (p), sedimentary (s), and volcanic (v). The relative proportions of Quaternary deposits in each catchment not only significantly vary, ranging from 4.5% (#6) to 48.3% (#2), but the prevailing type of Quaternary deposits also differs with dominant alluvial, colluvial, and glacial/fluvioglacial sediments in e.g. catchments #16, 11, and 14, respectively (Fig. 2.3). Further details on the lithological specificities of each group of catchments are given in Supplementary Material (A.1).

Latitudinal variations in catchment-averaged elevation, slope, k_{sn} , and annual precipitation range from 260 to 862 m, 8 to 21°, 8 to 69 m^{0.9}, and 748 to 1767 mm, respectively (Table 2.1 and Fig. 2.4). Whereas elevation, slope, and annual precipitation vary by a factor of 2 to 3 at the massif scale, a southward increasing trend in k_{sn} of almost one order of magnitude is documented. Mean elevation (Fig. 2.4A), slope (Fig. 2.4B), k_{sn} (Fig. 2.4C), and annual precipitation (Fig. 2.4D) all display a significant inverse correlation with the catchment-centroid latitude ($R^2=0.42-0.81$) and thereby attest the VM's N-S gradient. A secondary W-E gradient is also observed for topographic and climatic metrics, with two diverging trends. Unlike slope and k_{sn} values that are higher in the east than in the west (Fig. 2.4B-C), higher annual precipitation is recorded in the west (Fig. 2.4D). No significant W-E difference is found concerning the elevation (Fig. 2.4A).

2.4.2 ¹⁰Be and ²⁶Al concentrations and ²⁶Al/¹⁰Be ratios

Measured ¹⁰Be and ²⁶Al concentrations, calculated ²⁶Al/¹⁰Be ratios, and associated uncertainties are presented in Table 2.1. ¹⁰Be and ²⁶Al concentrations range from 6.1(±0.2) to 13.7(±0.5) × 10⁴ at/g and from 11.1(±2) to 92.5(±9) × 10⁴ at/g, respectively. ²⁶Al/¹⁰Be ratios for the 22 samples range from 1.4 ± 0.2 to 8.0 ± 0.7. ²⁶Al/¹⁰Be ratios are plotted in the two-nuclide diagram (Lal (1991); Fig. 2.5A) using CosmoCalc version 3.1 (Vermeesch, 2007) with the time-independent Lal (1991)/Stone (2000) scaling model. In this diagram, the steady-erosion island is comprised between two lines representing the evolution of the ²⁶Al/¹⁰Be ratio in material being continuously exposed without erosion (upper line) and varying

TABLE 2.1: Sample location, catchment characteristics, and cosmogenic nuclide results based on ^{10}Be and ^{26}Al measurements in quartz from stream sediment. ^aCatchment area associated with sample location, determined from a 25-m resolution DEM (BD ALTI). ^bMean elevation, ^cslope, and ^dnormalised channel steepness index (using the Mudd et al.'s 2018 LSD Topotools software with a reference concavity of 0.45) are all computed from the same DEM. ^eMean annual precipitation determined from the national hydroclimatic database of Météo-France (Vidal et al., 2010) over the 1958–2009 period. ^fGlacial imprint based on the percentage of the catchment being ice-covered during MIS2 according to Flageolet 2002. ^g> 10% : yes; < 10% : no. ^hMain lithology based on the hierarchical clustering analysis performed on bedrock lithologies and Quaternary deposits according to the BRGM's geological map (see Section 2.3.1). ⁱs : sedimentary, ^m : metamorphic, ^p : plutonic, ^v : volcanic. ^{h-i}Cosmogenic nuclide results for ^{10}Be and ^{26}Al . Concentrations are corrected for a blank contribution. Analytical uncertainties (1σ) include both the AMS counting statistics and internal error (0.5% for ^{10}Be ; Arnold et al. (2010)), chemical blank measurement and ^{27}Al measurements. Denudation rates are inferred using CosmoCalc version 3.1 (Vermeesch, 2007), based on the time-independent Lal (1991)/Stone (2000) scaling model. We used a density of 2.7 g/cm^3 and did not consider topographic shielding. Mean production rates were calculated from the latitude of the catchments centroids and their mean elevation. Muonic productions were computed following Braucher et al. (2011). Reported uncertainties correspond to the 1σ analytical errors. Integration time corresponds to the time spent in the equivalent mean attenuation length ($\sim 60\text{ cm}$). ^jMinimum burial time assuming a simple burial scenario for non-steady-state samples, calculated using CosmoCalc version 3.1 (Vermeesch, 2007).

River name	id	Latitude (°)	Longitude (°)	Area ^e (km ²)	Mean elevation ^b (m)	Mean slope ^c (°)	Mean k_{s0} (m ^{0.5})	Mean annual precipitation ^f (mm)	Glacial imprint ^f	Main lithology ^g	^{10}Be 10 ⁴ (at/g) qtz	[^{10}Be] denudation rate (mm/ka)	[^{10}Be] integration time (ka)	^{26}Al 10 ⁴ (at/g) qtz	[^{26}Al] denudation rate (mm/ka)	[^{26}Al] integration time (ka)	$^{26}\text{Al}/^{10}\text{Be}$	Minimum burial time ^j (ka)
Bruche	1	48.57	7.53	622.0	480.0	11.9	34.7	931.7	no	s	9.8 ± 0.3	46 ± 1	13.0	74 ± 6	45 ± 4	13.4	7.6 ± 0.7	/
Cleure	2	48.02	6.68	77.5	701.5	11.6	41.6	1357.6	yes	p	7.7 ± 0.5	70 ± 4	8.6	11.1 ± 2	363 ± 53	1.7	1.4 ± 0.2	3351.3
Doller	3	47.76	7.09	120.8	683.3	18.4	37.7	1394.2	yes	v	6.1 ± 0.2	88 ± 4	6.8	20.8 ± 4	189 ± 36	3.2	3.4 ± 0.7	1570.7
Fecht	4	48.12	7.33	246.0	707.1	18.2	54.6	1157.3	yes	p	10.5 ± 0.4	51 ± 2	11.7	63.8 ± 16	62 ± 16	9.7	6.1 ± 1.5	/
Fecht	5	48.04	7.13	91.4	862.4	20.4	65.8	1302.8	yes	m	11 ± 0.5	55 ± 2	10.9	67.3 ± 6	66 ± 6	9.1	6.1 ± 0.6	/
Giessen	6	48.30	7.36	121.4	471.3	15.9	29.9	1005.4	no	m	12.2 ± 0.4	36 ± 1	16.5	70.2 ± 5	47 ± 4	12.8	5.7 ± 0.5	501.2
Lauch	7	47.92	7.30	83.1	756.6	18.7	68.7	1095.2	yes	m	8.5 ± 0.3	66 ± 2	9.1	56.5 ± 5	73 ± 7	8.3	6.6 ± 0.6	/
Liepvrette	8	48.28	7.36	127.8	557.6	20.0	37.3	1105.8	no	m	9.1 ± 0.3	53 ± 2	11.4	61.6 ± 5	57 ± 5	10.5	6.8 ± 0.6	/
Meurthe	9	48.52	6.60	1035.0	508.4	11.2	24.5	1224.5	no	s	8.7 ± 0.4	53 ± 3	11.3	69.5 ± 5	49 ± 4	12.3	8 ± 0.7	/
Modre	10	48.86	7.51	132.4	260.2	8.3	8.0	788.0	no	s	7.3 ± 0.4	52 ± 3	11.5	51.5 ± 6	55 ± 6	10.9	7 ± 0.9	/
Mosig	11	48.59	7.48	148.8	358.8	8.4	24.7	747.8	no	s	12 ± 0.4	34 ± 1	17.6	73 ± 6	41 ± 3	14.5	6.1 ± 0.5	/
Moselle	12	48.12	6.54	1037.0	681.0	12.6	32.5	1333.3	yes	p	9.8 ± 0.3	54 ± 2	11.2	38.1 ± 3	103 ± 9	5.8	3.9 ± 0.4	1310.3
Moselotte	13	48.02	6.69	258.4	789.7	15.4	46.0	1357.0	yes	p	7.1 ± 0.4	81 ± 4	7.4	34.9 ± 4	121 ± 16	4.9	4.9 ± 0.7	832.5
Ognon	14	47.76	6.61	96.9	634.8	15.4	35.3	1545.6	yes	v	9.3 ± 1.3	55 ± 8	11.0	14.2 ± 6	267 ± 110	2.2	1.5 ± 0.7	3198.7
Plaine	15	48.41	6.85	126.8	492.5	15.8	20.0	1180.7	no	s	10.2 ± 0.3	45 ± 2	13.4	75.9 ± 6	44 ± 4	13.6	7.4 ± 0.6	/
Sarre	16	48.69	7.02	188.8	437.4	11.8	20.0	1043.7	no	s	9.9 ± 0.3	44 ± 1	13.6	63 ± 6	51 ± 5	11.7	6.3 ± 0.6	/
Savoureuse	17	47.68	6.83	62.4	647.2	13.4	25.8	1766.9	yes	v	8.3 ± 0.3	62 ± 2	9.7	55.9 ± 7	67 ± 8	8.9	6.8 ± 0.8	/
Thur	18	47.80	7.16	210.6	717.1	21.3	39.5	1256.8	yes	m	6.8 ± 0.3	80 ± 3	7.5	35.6 ± 4	113 ± 12	5.3	5.2 ± 0.6	701.4
Vologne	19	48.15	6.63	286.8	652.6	9.7	25.3	1338.1	yes	p	11.5 ± 0.6	45 ± 2	13.4	46.6 ± 4	82 ± 8	7.3	4.1 ± 0.4	1210.6
Weiss	20	48.13	7.33	166.1	709.7	16.9	65.6	1033.0	no	p	13.7 ± 0.5	39 ± 1	15.3	92.5 ± 9	42 ± 4	14.2	6.7 ± 0.7	/
Zinsel du Sud	21	48.80	7.38	119.8	293.7	9.5	9.6	807.4	no	s	6.2 ± 0.2	64 ± 3	9.4	42.1 ± 4	69 ± 7	8.7	6.8 ± 0.8	/
Zorn	22	48.76	7.45	400.5	337.9	10.8	17.5	860.1	no	s	8.6 ± 0.3	47 ± 2	12.8	49.7 ± 4	61 ± 4	9.9	5.7 ± 0.5	508.7

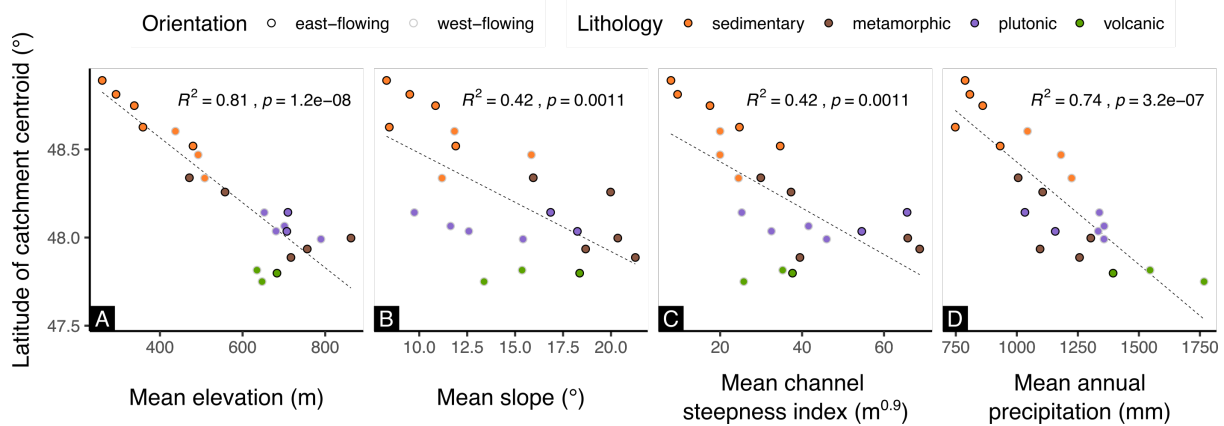


FIGURE 2.4: Significant relationships between the latitude of catchment centroids and the morphometric/climatic factors in the VM. (A) Mean elevation. (B) Mean slope. (C) Mean normalized channel steepness index. (D) Mean annual precipitation. A second spatial gradient is also observed between the eastern- and the western-flowing catchments. Note that statistics (R^2 ; p-value) are calculated using total least square regressions since the latitude does not depend on the controlling factors.

steady erosion rates (lower line). Thirteen samples (#1, 4, 5, 7, 8, 9, 10, 11, 15, 16, 17, 20, 21) out of 22 modern stream samples ($\sim 60\%$) plot within the steady-erosion island, considering 2σ uncertainties (Fig. 2.5A). They are referred to as the steady-state samples hereinafter. The remaining nine samples (#2, 3, 6, 12, 13, 14, 18, 19, 22) plot below the steady-erosion island, considering 2σ uncertainties (Fig. 2.5A). In the so-called complex exposure zone, these samples show different apparent burial durations (i.e., minimum burial times) due to their varying depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios (Fig. 2.5A): these range from ~ 500 ka (#6, 22) to more than 3 Ma (#2, 14; Table 2.1). They are referred to as the non-steady-state samples hereinafter.

2.4.3 Catchment-averaged denudation rates

We consider here the whole dataset regardless of the individual $^{26}\text{Al}/^{10}\text{Be}$ ratios (Table 2.1 and Fig. 2.5B). CWD rates and integration times inferred from ^{10}Be concentrations range from 34 ± 1 to 88 ± 4 mm/ka (catchments #11; 3) and 6.8 to 17.6 ka, respectively. Those inferred from ^{26}Al concentrations range from 41 ± 3 to 363 ± 53 mm/ka (catchments #11; 2) and 1.7 to 14.5 ka, respectively (Table 2.1 and Fig. 2.5B). First, when comparing ^{10}Be and ^{26}Al -inferred CWD rates (Fig. 2.5B), the steady-state samples are well distributed along the 1:1 line ($\pm 25\%$). By contrast, among the nine non-steady-state samples, seven of them (#2, 3, 12, 13, 14, 18, 19) clearly plot apart from the 1:1 line (Fig. 2.5B). Almost all ^{26}Al -inferred CWD rates are significantly higher than ^{10}Be -inferred values, sometimes by a factor four or higher (#2 and 14). Second, we plot the latitude of the catchments' centroid against CWD rates inferred from both nuclides, alongside the four lithological groups and the mean annual precipitation (Fig. 2.6). On the eastern side, a southward increasing trend is highlighted for both ^{10}Be and ^{26}Al -derived CWD rates. On the western side, no clear spatial pattern can be observed,

neither for ^{10}Be nor for ^{26}Al . Importantly, the seven non-steady-state samples showing the largest discrepancies in ^{10}Be - versus ^{26}Al -CWD rates are all located in the southern part of the VM (Fig. 2.6), regardless the catchment lithology and annual precipitation.

2.5 Discussion

2.5.1 What can - and cannot - be said about spatial denudation patterns in the Vosges Mountains?

2.5.1.1 Unsteadiness blurs cosmogenic nuclide signals at the massif scale

A global southward increasing trend in elevation, slope, k_{sn} and annual precipitation is highlighted throughout the VM together with a clear latitudinal distribution of the main bedrock lithologies (Figs. 2.3,2.4). These have been proposed as the primary factors governing catchment-wide denudation measured in active stream sediments (e.g., Safran et al., 2005; Carretier et al., 2013; Cyr et al., 2014; Mandal et al., 2015; Olen et al., 2016; Olivetti et al., 2016; Delunel et al., 2020; Peifer et al., 2021). Correlations are thus tested for elevation, slope, k_{sn} and annual precipitation (Fig. 2.7; Table A.1 in Supplementary Material). Only one significant relationship (though weak: $R^2 = 0.22 - 0.24$; $p\text{-value} < 0.03$) is observed between annual precipitation and CWD rates inferred from both nuclides and based on the whole VM dataset (Fig. 2.7B, E).

One must recall, however, that nine catchments (#2, 3, 6, 12, 13, 14, 18, 19, 22; $\sim 40\%$) violate the steady-state assumption required to infer denudation rates from measurement of *in situ* cosmogenic nuclides concentrations in river sediments (von Blanckenburg, 2005; Dunai, 2010). Accordingly, these non-steady-state samples should be discarded when it comes to (i) infer CWD rates at the massif scale and (ii) correlate these with potential controlling factors (climatic and/or morphometric control). Discarding these catchments has a strong impact as they represent a total drainage area of $\sim 1,987 \text{ km}^2$ (i.e. 43% of the study area; Fig. 2.1). Moreover, these catchments are mostly located in the VM's southern part (7/9) and thereby substantially reduce the relationship between the VM's N-S structuration and the spatial pattern of denudation (Fig. 2.7). Unsteadiness in cosmogenic nuclide concentrations measured in modern river sediments from nine catchments thus prevents any conclusions regarding the controlling factors at the whole massif scale.

Two subsets can be distinguished among non-steady-state samples: the northern ones (# 6 and 22) have slightly depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios (5.7 ± 0.5) whereas the southern ones (#2, 3, 12, 13, 14, 18, 19) have much lower ratios ($1.4 \pm 0.2 - 5.2 \pm 0.6$; Fig. 2.5A; Fig. 2.6). Whilst ^{10}Be - and ^{26}Al -inferred CWD rates of the first group overlap the 25% uncertainty space around

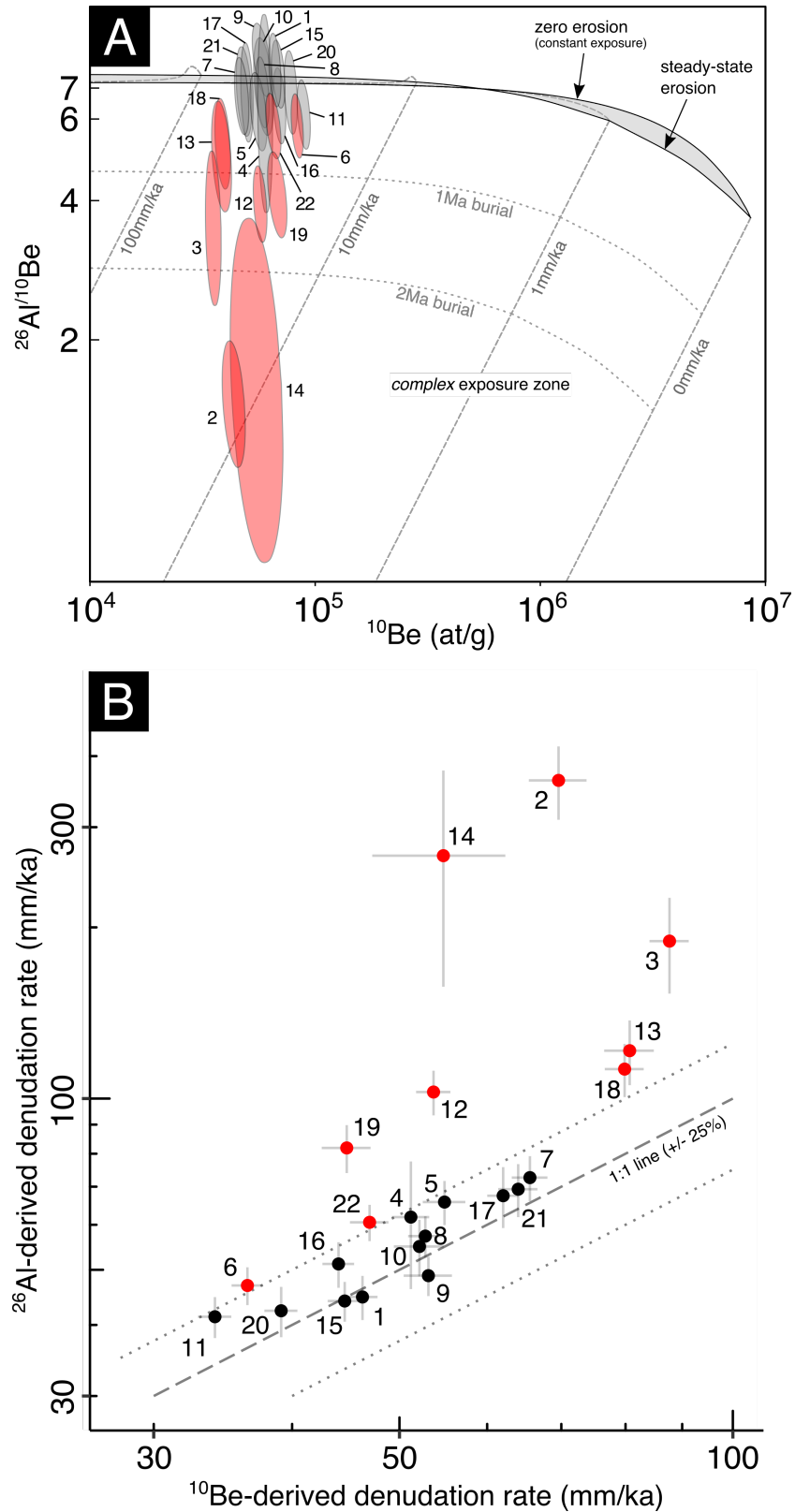


FIGURE 2.5: A: Two-nuclide diagram plotting $^{26}\text{Al}/^{10}\text{Be}$ ratios versus ^{10}Be concentrations. Sample ellipses correspond to 2σ analytical errors. Thirteen samples plot within the steady-erosion island and thus experienced steady-state erosion (gray ellipses). Nine samples plot below the steady-erosion island and thus experienced complex exposure histories (red ellipses). This plot was produced using CosmoCalc version 3.1 (Vermeesch, 2007), with the time-independent Lal (1991)/Stone (2000) scaling model. Muonic production was based on Braucher et al. (2011). B: Distribution of ^{26}Al -derived versus ^{10}Be -derived CWD rates. Error bars correspond to the 1σ analytical errors. Red and black points correspond to the non-steady-state and steady-state catchments, respectively (panel A). Note the logarithmic scales for both plots.

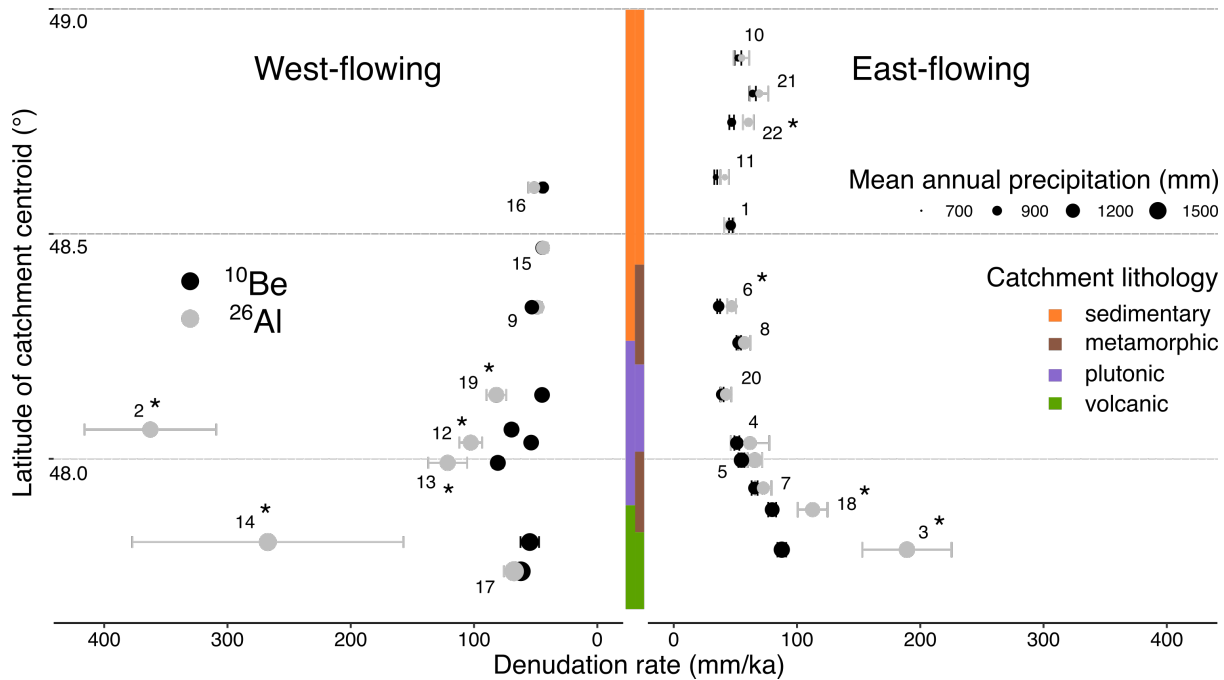


FIGURE 2.6: North-South distribution of CWD rates inferred from ^{10}Be (black points) and ^{26}Al (gray points) concentrations, along with the main lithologies and mean annual precipitation of the studied catchments. Horizontal error bars represent the 1σ uncertainty associated with the inferred CWD rates. Non-steady-state catchments are marked with an asterisk. The seven non-steady-state samples delivering the largest discrepancies in denudation values are all located in the southern part of the VM, regardless of the catchment's lithology.

the 1:1 line defined by both nuclides (Fig. 2.5B), those of the second group systematically plot apart from it. We assume that the first two samples (#6 and 22) might possibly fall into the range of uncertainties related to methodological effects on depletion of $^{26}\text{Al}/^{10}\text{Be}$ production ratios (Wittmann et al., 2020). On the other hand, the strongly depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios peculiar to the second group cannot be explained by intrinsic uncertainties linked to the production ratio or attenuation lengths. One must consider instead the effect(s) of one or several geomorphological processes since no significant relationship is observed between catchment area and $^{26}\text{Al}/^{10}\text{Be}$ ratios at the massif scale (Supplementary Material A.2, Table A.1, Fig. A.3).

2.5.1.2 Lithological and vegetation potential control on denudation for steady-state catchments

With removing the nine non-steady-state samples, the range of ^{10}Be - and ^{26}Al -derived CWD rates in the thirteen steady-state catchments is significantly reduced to $34 \pm 1 - 66 \pm 2$ and $41 \pm 3 - 73 \pm 7$ mm/ka, respectively. ^{10}Be -derived CWD from a small unglaciated catchment located between our catchments #8 and 20 fall in the same range (i.e. ~ 38 mm/ka; Ackerer et al., 2022). By contrast, in these thirteen catchments, the observed large ranges in mean elevation, k_{sn} and annual precipitation remains as in the initial dataset (see 2.4.1) and only

the slope fractionally decreases (between 8.3 and 20.4°; Table 2.1). Yet, none of the above morphometric and climatic controlling factors appears playing a major role here, as shown by the absence of significant correlation with CWD rates inferred from both nuclides (Supplementary Material A.2, Table A.1, Fig. A.2).

The VM denudation pattern, including the previously-reported lack of correlation with morphometric/climatic parameters, can be tentatively explained by two main features peculiar to this low-elevation mid-latitude mountain range. First, deciduous, resinous and mixed forests currently cover $\sim 70\%$ of its surface (Rochel, 2004), mostly on hillslopes and inter-fluve areas (Fig. 2.2B). They were probably even more widespread before the Middle Age as forest clearance linked to past agricultural practices occurred at that time (Keller, 2023). Whilst vegetation has been increasingly considered as a controlling or modulating factor on CWD rates (Mandal et al., 2015; Olen et al., 2016; Schaller et al., 2018), Acosta et al. (2015) reported lower denudation rates in areas covered by dense vegetation. We likewise suggest that, in addition to this possible denudation lowering effect from dense vegetation, the VM's dense forest cover could have smoothed the variability in topographic surface erodibility. This would have resulted in lower amplitude of CWD rates that do not much differ across the thirteen investigated catchments, despite the contrasted ranges in elevation, slope, k_{sn} and annual precipitation.

Second, and more importantly, the lithological heterogeneity encountered within the thirteen steady-state catchments possibly overprints the other controlling factors (Morel et al., 2003; Campforts et al., 2020). In the following we compare the crystalline- (#4, 5, 7, 8 and 20) and sedimentary-dominated (#1, 9, 10, 15, 16 and 21) catchments (Figs. 2.3 and 2.6). Although mean k_{sn} value of the crystalline group ($\sim 58 \text{ m}^{0.9}$) is about three times higher than that of the sedimentary group ($\sim 19 \text{ m}^{0.9}$), ^{10}Be - and ^{26}Al -derived CWD do not significantly differ between both groups: $\sim 53/ \sim 60 \text{ mm/ka}$ versus $\sim 51/ \sim 52 \text{ mm/ka}$ in crystalline and sedimentary catchments, respectively (Supplementary Material A.2, Fig. A.2). Despite an observed lower magnitude in relief incision (i.e. $\sim 150 \text{ m}$) in the sedimentary-dominated landscapes of the northern VM, denudation there is assumed to be as high as in the much more incised southern part (Figs. 2.1C and 2.2). The observed similarity in CWD rates obtained for both lithological groups thus implies that the erodibility of the homogeneous cover of Triassic sandstones should be much higher than that of Palaeozoic granites and gneisses. Note that all these catchments are located at the margins of the former glacier extents and, accordingly, must have experienced periglacial conditions during Quaternary cold stages. In addition, sandstone's enhanced erodibility due to periglacial processes has been already recognised in Palaeozoic, European low-mountain ranges (e.g. Ardennes in Belgium, Rixhon, 2010) and was recently quantified by cosmogenic nuclides in sandstone-dominated catchments for the central Appalachians (Del Vecchio et al., 2022).

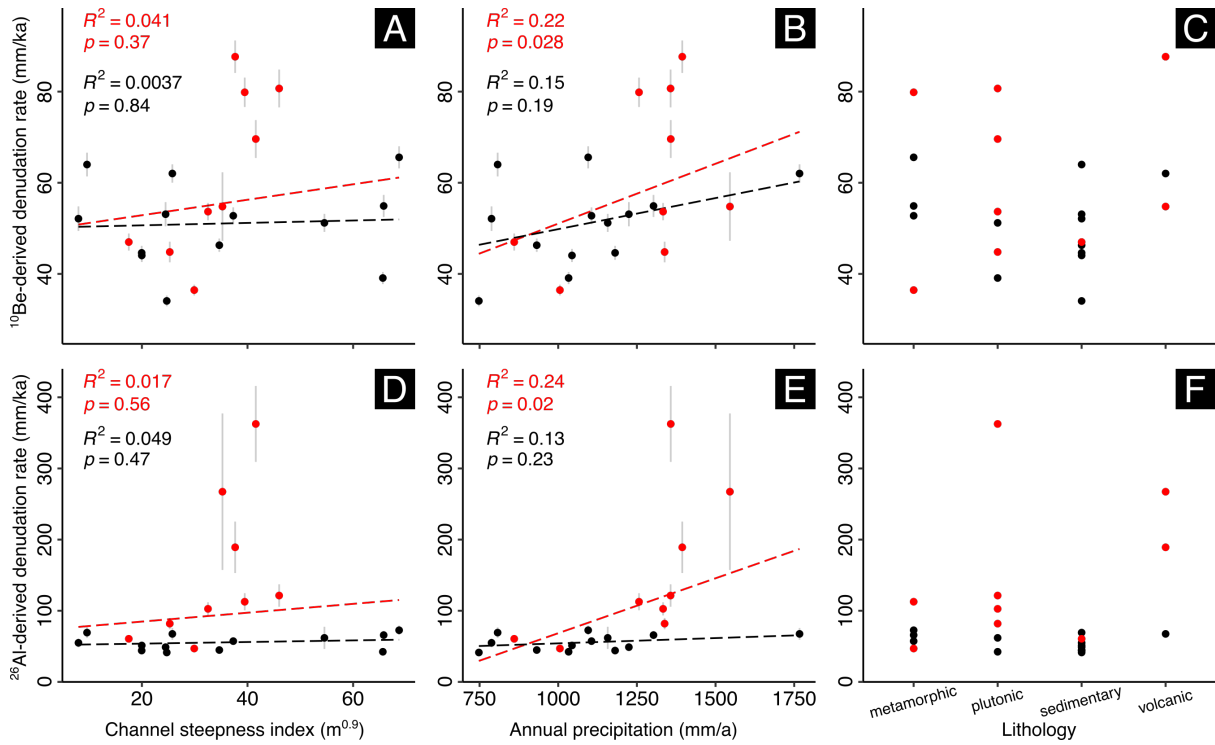


FIGURE 2.7: Relationships between CWD rates inferred from ^{10}Be (A-C) or ^{26}Al (D-F) concentrations and potential controlling factors: normalized channel steepness index, annual precipitation, and main lithologies. (Red) black points correspond to the (non)-steady-state samples. Statistics are computed from linear regressions using the whole dataset (red lines) or the steady-state subset (black lines). A weak but significant relationship is observed between CWD rates and mean annual precipitation (p -value < 0.03). The significance of the relationship is lost when excluding the non-steady-state samples from the dataset (black lines).

2.5.2 Understanding the impacts of former glaciations on cosmogenic nuclide disequilibrium

One of the main outcomes of this study is the spatial correlation between catchments in cosmogenic disequilibrium and the reconstructed former glaciations covering the southern VM. Our aim here is to assess the impact of glaciations on the depletion of the $^{26}\text{Al}/^{10}\text{Be}$ ratio through two processes, glacial shielding and reworking of previously buried sediment. Among all processes listed by Wittmann et al. (2020) resulting in depleted Al/Be ratio in river sediments, these two are the only ones that are relevant in the VM.

2.5.2.1 $^{26}\text{Al}/^{10}\text{Be}$ ratio's depletion correlates with repeated glacial shielding

Although glacial shielding is one of the best documented processes to cause $^{26}\text{Al}/^{10}\text{Be}$ -ratio lowering in bedrock of high-latitude, formerly glaciated regions (Corbett et al., 2013; Strunk et al., 2017; Andersen et al., 2018, 2020), it was barely considered when assessing CWD rates from modern stream sediments. In the following, we show a quantitative comparison between

$^{26}\text{Al}/^{10}\text{Be}$ ratios in modern river sediments and former glacial covers in each catchment attributed to the LGM/MIS2 (Flageollet, 2002) and the penultimate glaciation/MIS6 (Flageollet, 2002) because their reconstruction is based on reliable and widespread field evidence.

Main outputs of this comparison are twofold. First, significant correlations (p -values < 0.0001) are observed between $^{26}\text{Al}/^{10}\text{Be}$ ratio and the extent of past glacial cover during MIS2 ($R^2=0.53$) and 6 ($R^2=0.62$; Fig. 2.8A-B). All unglaciated catchments (i.e. $< 10\%$ of ice cover during MIS2) cluster on the 6.7 production-ratio line, regardless the considered glacial period. The eleven formerly ice-covered catchments show an inverse general trend for both cold periods: the larger the former glacial cover in each catchment, the lower the $^{26}\text{Al}/^{10}\text{Be}$ ratio measured in its stream sediments (Flageollet, 2002). This observation underlines the probable link between long-lasting (i.e. 10^{3-4} years) and repetitive ice shielding and unsteadiness in the stream cosmogenic nuclide signal. One must bear in mind here that pre-MIS6 glaciations were documented in the VM's southern part (Flageollet, 2002) and suggested at its southern margin (Ménillet, 2019). We thus propose that shielding in this area probably occurred during many cold periods of the Quaternary even if the correlated deposits of these pre-MIS6 glaciations can only be found in fluvial terrace sequences (e.g. Moselle and Ognon) located at the massif's outskirt (Harmand, 2020). Second, the significance of these relationships increases with larger ice shielding between the two glacial periods: the ice-covered area roughly doubles between MIS2 ($\sim 730 \text{ km}^2$) and 6 ($\sim 1570 \text{ km}^2$; Fig. 2.1E). Note that three catchments were fully ice-covered during the LGM (i.e. $> 95\%$; Fig. 2.8B). We suggest here that this R^2 increase (0.53 to 0.62) is probably linked to distinct ice dynamics reflected in the spatial pattern peculiar to each cold stage (Flageollet, 2002). The LGM was characterised by a mosaic of valley glaciers with various ice-flow directions and lengths (up to $\sim 40 \text{ km}$ in the Moselle, #12), leaving large parts of the VM's main divide and secondary ridges ice-free (Flageollet, 2002). By contrast, only a handful of small-dimensioned nunataks protruded the almost continuous MIS6's ice field characterised by a former SW-NE extent of $\sim 60 \text{ km}$ (Flageollet, 2002). Although reconstructing the thermal nature of the former glacial masses in the southern VM is beyond the scope of this study, two lines of arguments based on previous studies using ^{10}Be and ^{26}Al measurements could point to limited subglacial erosion and thus previous cold-based ice cover. First, if occurrence of cold-based glaciers is traditionally assigned to arctic regions (Copland, 2013), the highest ridges in the northern Appalachians, i.e. a similar setting to the southern VM, were covered by cold-based ice during the LGM (Bierman et al., 2015). Second and most important, depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios in bedrock and erratics from Greenland were well explained by prolonged and repeated burial beneath cold-based ice during glaciations (Corbett et al., 2013). The same holds for glacial tills shielded by cold-based ice on the Baffin Island, where the sampled grain size matches the sandy fraction studied here (Staiger et al., 2006). The depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios of these eight sandy samples (i.e. 1.3 to 4.3) fall into a similar range to ours (1.4 to 5.2). Such depleted values would imply

minimum burial times ranging from ~ 500 ka to ~ 3 Ma (Fig. 2.5; Table 2.1). If this continuous burial duration was indeed observed on Baffin Island, it is not so in the VM, whose relief has certainly been re-exposed to radiation during each interglacial period. It therefore seems highly probable that VM sediments were not correctly re-exposed to cosmic-rays between glaciations. Knudsen and Egholm (2018) modelling results showed that very low $^{26}\text{Al}/^{10}\text{Be}$ ratios may not be exclusively explained by prolonged burial under non-erosive ice, but may also reflect spatio-temporally varying bedrock exposure/burial and erosion histories. In addition to ice shielding, we therefore suggest that repeated, complex sediment routing systems during deglaciation and subsequent interglacial periods probably characterised the formerly glaciated catchments of the VM, thereby explaining the depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios measured in the stream sediments.

2.5.2.2 $^{26}\text{Al}/^{10}\text{Be}$ ratio's depletion correlates with the amount of glacial sediment storage

Successive sediment burial and reworking within a sediment routing system, sometimes for long periods of time (i.e. $> 10^5$ years; Fülöp et al., 2020; Bhattacharjee et al., 2023) or rapid changes in denudation rates may yield to depleted $^{26}\text{Al}/^{10}\text{Be}$ ratio measured in the stream sediments. Although the cosmogenic nuclide signal is likely to be preserved from the source area to the sink in most cases (Wittmann and von Blanckenburg, 2009), a handful of studies highlighted a significant decrease in $^{26}\text{Al}/^{10}\text{Be}$ ratios measured in stream sediments from glaciated catchments (Safran et al., 2005; Hippe et al., 2012; Grin et al., 2018; Wittmann et al., 2020; Zhang et al., 2021). Glacial environments are generally well-known to have complex sediment routing systems (Porter et al., 2019). Based on CWD rates inferred from both nuclides, Zhang et al. (2021) also identified this issue but did not provide further explanations regarding the correlated processes. This study presents, to our knowledge, the first quantitative comparison between $^{26}\text{Al}/^{10}\text{Be}$ ratios in modern river sediments and the surficial proportions of Quaternary deposits at the individual catchment scale (Flageollet, 2002).

When all Quaternary sediments are considered, regardless of their nature, a significant correlation is observed ($R^2=0.33$; $p\text{-value}<0.01$; Fig. 2.8C). When only glacial/fluvio-glacial sediments are considered, the strongest correlation is observed among all our proxies to assess glacial impact on measured $^{26}\text{Al}/^{10}\text{Be}$ ratios ($R^2=0.67$; $p\text{-value}<0.0001$; Fig. 2.8D), including ice-cover shielding. Whereas all unglaciated catchments cluster on the 6.7 production-ratio line, higher depletions of $^{26}\text{Al}/^{10}\text{Be}$ ratios are observed with increasing percentage of glacial sediments, peaking in catchment #2 (Fig. 2.8D). This suggests that (i) glacially-reworked deposits contain more $^{26}\text{Al}/^{10}\text{Be}$ -depleted material than fluvial and colluvial ones, and (ii) the greater the amount of glacial deposits present in the catchment, the greater the likelihood for the current river channel to rework these deposits.

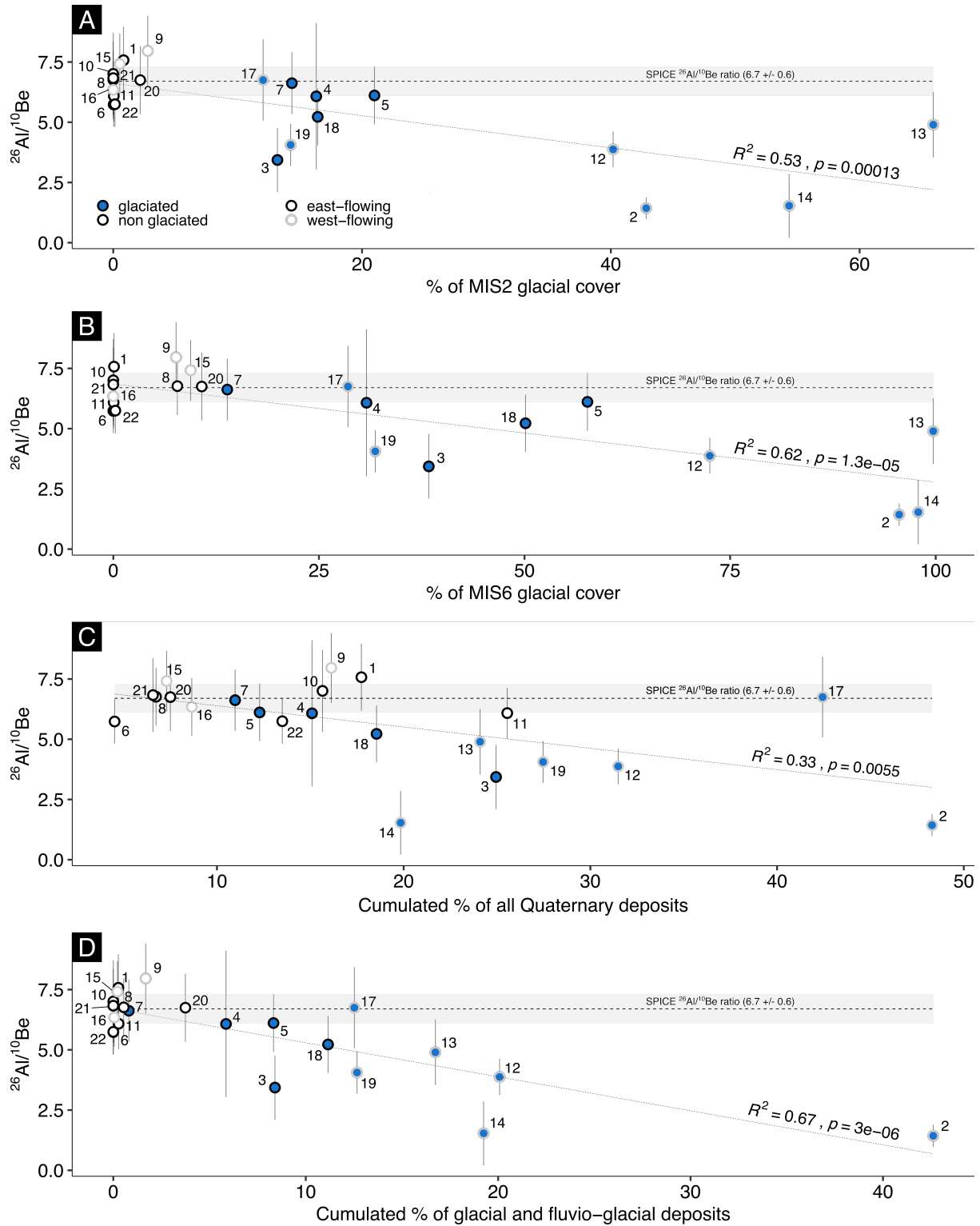


FIGURE 2.8: Relationships between $^{26}\text{Al}/^{10}\text{Be}$ ratios and (A) percentage of the catchment being ice-covered during MIS2, (B) same for MIS6, (C) cumulated percentage of all Quaternary deposits (i.e., alluvial, colluvial, glacial, and fluvio-glacial deposits), and (D) cumulated percentage of glacial and fluvio-glacial deposits. Glacial covers (A and B) are extracted from Flageollet's 2002 mapping. Quaternary sediment covers (C and D) are extracted from the BRGM's geological map. The vertical error bars correspond to the $^{26}\text{Al}/^{10}\text{Be}$ ratios at 1σ uncertainty. Blue points highlight the glaciated catchments (i.e. >10% of ice cover during MIS2). The horizontal dashed line corresponds to the surface production ratio recently determined by the SPICE project (Fenton et al., 2022). We did not use the surface production ratio from Cosmocalc (Vermeesch, 2007) as it does not provide any uncertainty. The strongest significant linear regression ($R^2 = 0.67$, $p\text{-value} = 3 \times 10^{-6}$) is observed with the cumulated percentage of glacial and fluvio-glacial deposits (D).

A closer look at the spatial distribution of glacial/fluvio-glacial deposits throughout the southern VM clearly highlights a W-E gradient, with the highest proportions of glacial/fluvio-glacial sediments encountered in the west-flowing catchments (i.e. #2, 12, 13, 14, 17, 19; Fig. 2.8D and Supplementary Material Fig. A.11). One must recall here the geologically controlled W-E increasing gradient in mean catchment slope: 12.6 vs 16.9° to the west and the east of the main divide, respectively. Efficient sediment transfer in the east-flowing catchments can be assumed because of (i) shorter transport distances (Fig. 2.1B), (ii) higher mean slopes (Fig. 2.4B), (iii) much lower proportions of glacial deposits along their intra-massif reaches (Supplementary Material Fig. A.11) and (iv) km-large, thick, Quaternary alluvial fans extending at the eastern Vosgian footslopes (e.g. catchments #1, 4, 6, 8, 18; Wuscher, 2021). As a result, only one of the east-flowing catchments (#3) plots apart from the 6.7 line (Fig. 2.8).

By contrast, among the six west-flowing, formerly glaciated catchments, five of them show depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios (Fig. 2.8). This probably reveals more complex sediment routing systems and the catchment #2 (i.e. the so-called Cleurie) with the most depleted ratio of our whole dataset (~ 1.4) epitomises this. First, if this catchment belongs to the group of catchments that were significantly and repeatedly covered by glaciers (#2, 12, 13, 14; Fig. 2.8A-B), it clearly stands out because of the highest relative proportion of glacial/fluvio-glacial sediments (i.e. $>40\%$ vs $<20\%$ for the rest; Fig. 2.8D). Though being the second smallest catchment investigated in this study ($<80\text{ km}^2$; Table 2.1), the Cleurie accordingly hosts the highest density and largest diversity of glacial and glacial-related landforms, including e.g. end-moraines, kame terraces, lacustrine and deltaic deposits (Seret, 1967; Flageollet, 2002). Second, thicknesses of valley-floor sediments exceeding 50 m, though undated, were revealed by drillings performed in the uppermost part of the catchment (BRGM). Third, petrographical and mineralogical analyses performed in some of these deposits reveal a partially allochthonous origin for some of the sampled materials, i.e. from the catchment located to the south (#13; Flageollet et al. (1971)). Fourth, the peculiar spatial distribution and geometry of the best-preserved, i.e. tens of meters high, end moraines strongly point to a triple convergence of valley glaciers supposedly occurring during the LGM and peaking in the formation of a palaeolake in the catchment's upstream part (Flageollet, 2002). Fifth, a major flood occurred in the valley in 1770, whose damages were mapped in an archive document (Salmon, 1770). The latter reveals erosion and reworking of glacial material buried at the base of the highest moraines. Altogether, the concentration and diversity of glacial(-related) landforms along with the long-lasting storage of reworked and partially allochthonous material (i.e. "stop-and-go", possibly similar to foreland glacial systems, e.g. Coge et al., 2018) suggests a very complex sediment routing system in the Cleurie which yields the most depleted $^{26}\text{Al}/^{10}\text{Be}$ ratio of our dataset. More generally, our findings suggest that the sediments sampled in the SW catchments of the VM could have survived several glacial cycles. They might have been shielded by cold-based ice during glaciations, without being sufficiently re-exposed to cosmic rays during

interglacial periods as they were buried underneath thick glacial sediments. Measuring low $^{26}\text{Al}/^{10}\text{Be}$ ratios in modern stream sediment also implies that the river network is currently partly supplied by depleted $^{26}\text{Al}/^{10}\text{Be}$ sedimentary material.

2.5.3 Regional denudation in the Vosges and Black Forest Mountains

At the regional scale, we compare our CWD rates with those previously obtained in SW Germany in the neighbouring Black Forest Mountains which show a similar overall N-S structuration in lithology, elevation, and morphometry. We recalculated fifty ^{10}Be -inferred CWD rates published by Morel et al. (2003), Meyer et al. (2010), and Glotzbach et al. (2013) using the methodology reported above to allow quantitative comparisons (see Supplementary Material A.1). All sampled catchments are located within a similar latitude range to the VM (Fig. 2.9); they can thus be compared in terms of lithology, elevation, and morphometry. The median (and standard deviation) ^{10}Be -derived CWD rates for the Black Forest Mountains and VM, regardless of the cosmogenic nuclide steadiness of the samples, appear at first glance remarkably similar: 52.6 ± 23.6 and 52.9 ± 14.4 mm/ka, respectively (Fig. 2.9).

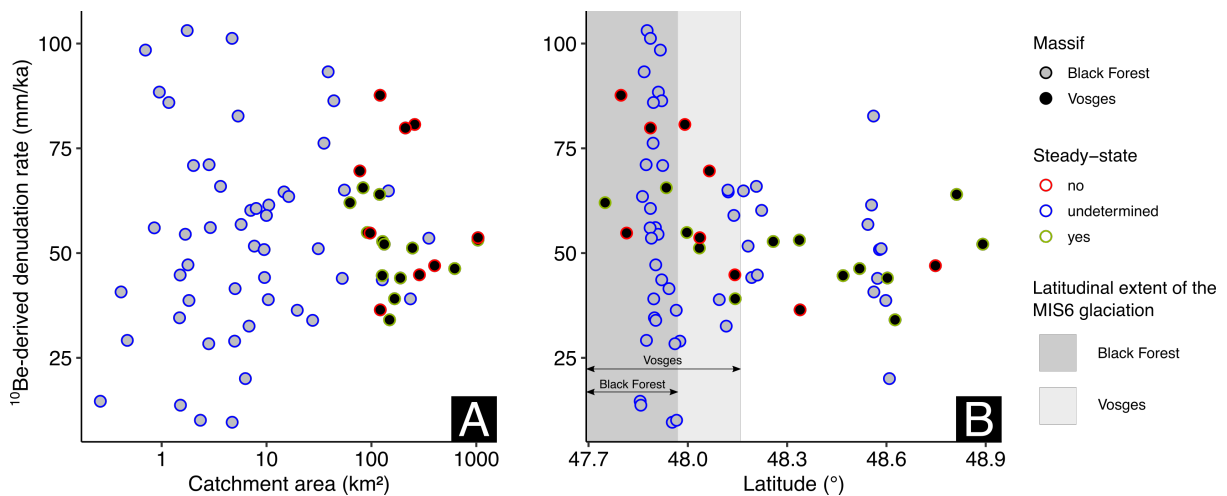


FIGURE 2.9: Recalculated ^{10}Be -derived CWD rates in the Black Forest, compared to the ^{10}Be -derived CWD rates in the Vosges Massif, vs. (A) catchment area and (B) latitude of the catchment's centroid. The dark and light gray zones correspond to the latitudinal MIS6 glaciation extent of the Black Forest and the Vosges Massif, according to Flageollet (2002) and Ehlers et al. (2011), respectively.

However, two complications considerably hamper a meaningful comparison between the two datasets. First, the average drainage area of the Black Forest Mountains catchments ($\sim 27 \text{ km}^2$) is one order of magnitude smaller than for the VM catchments ($\sim 262 \text{ km}^2$; Fig. 2.9A). Moreover, a majority of these small catchments are located in the headwaters of three main rivers draining the Black Forest Mountains, i.e. the Danube, Kinzig, and Wutach. This strongly contrasts with our sampling strategy consisting of sampling all main rivers draining the VM. Second and most importantly, almost all catchments sampled by Morel et al. (2003) and Glotzbach et al. (2013) are located within the former MIS6 glacier extent (specific sampling

approach to investigate the impact of glacial erosion and ice cover on CWD rates). However, the sole ^{10}Be measurement in modern stream sediments in these studies leaves this twofold fundamental question open: considering a cosmogenic nuclide non-steady state (Glotzbach et al., 2013), how much disequilibrium characterizes the sampled material there?

2.6 Conclusions and perspectives

From the massif-scale characterisation coupled to the paired-nuclide analysis to infer ^{10}Be - and ^{26}Al -based catchment-wide denudation rates across the Vosges Massif, we can conclude that:

1. elevation, slope, k_{sn} and precipitation are primarily organised along a N-S increasing trend while a subordinated E-W gradient occurs;
2. ^{10}Be - and ^{26}Al -derived catchment-wide denudation rates range from 34 ± 1 to 66 ± 2 , and 41 ± 3 to 73 ± 7 mm/ka, respectively, in thirteen steady-state catchments that are in cosmogenic equilibrium;
3. nine samples have depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios (up to 1.4) and they mostly cluster in the Vosges Massif's area (SW) which was the most intensively glaciated during Quaternary cold stages. They reveal cosmogenic nuclide unsteadiness in modern river sediments and must be accordingly discarded when it comes to infer catchment-wide denudation rates, thereby precluding any conclusion about controlling factors at the whole massif scale;
4. no relationship is reported between elevation, slope, k_{sn} and precipitation, and catchment-wide denudation rates in the steady-state catchments. Instead, lithological contrasts may control the pattern of denudation with a higher erodibility, especially under periglacial conditions, of the sandstone-dominated catchment compared to the crystalline-dominated catchments. We also suggest that the dense and almost ubiquitous vegetation cover reduces the variability in erodibility;
5. a statistically significant relationship is observed between the degree of depletion of the $^{26}\text{Al}/^{10}\text{Be}$ ratios and the spatial pattern of glaciation: the larger the former glacial cover in each catchment, the lower the $^{26}\text{Al}/^{10}\text{Be}$ ratio measured in its stream sediments. It underlines the link between long-lasting and repetitive ice shielding and unsteadiness in the stream cosmogenic signal. It also possibly suggests the past existence of cold-based glaciers in the Vosges Massif.
6. an even more significant relationship is reported between the degree of depletion of the $^{26}\text{Al}/^{10}\text{Be}$ ratios and the increasing proportion of glacial and fluvio-glacial deposits

within each catchment. It underlines the link between complex sediment routing systems in glacial environments and unsteadiness in the stream cosmogenic signal. This also well aligns with an E-W gradient: efficient sediment transfer versus longer sediment storage in the east- and west-flowing catchments, respectively. The catchment with the lowest $^{26}\text{Al}/^{10}\text{Be}$ ratio (1.4) remarkably matches the most complex sediment routing system from all sampled catchments.

This study raises several key points. First, we argue for systematic measurements of ^{26}Al to complement ^{10}Be and to test the steady-state assumption when it comes to infer catchment-wide denudation rates from modern stream sediments, especially in formerly glaciated landscapes. Second, we show that 86% of the catchments containing more than 10% of glacial/fluvio-glacial deposits (6/7) have depleted $^{26}\text{Al}/^{10}\text{Be}$ ratio and are thus in cosmogenic nuclide disequilibrium. A good knowledge of the regional glacial history, including quantitative and chronological data, is thus prerequisite and this 10% occurrence of glacial/fluvio-glacial deposits could be used, at first glance, as a threshold value. Third, despite strong quantitative evidence of $^{26}\text{Al}/^{10}\text{Be}$ ratio depletion through glacial related processes, further work is required to improve our interpretations. As for the impact of former ice covers, one should refine their spatial extents, assess their palaeothicknesses and chronologies. Abundant erratics and moraine deposits are key landforms here. ^{10}Be and ^{26}Al measurements of glacially polished bedrock should also be performed to assess glacial erosion efficiency and test the assumption of former cold-based glaciers. As for the impact of complex sediment-routing system, along stream ^{10}Be and ^{26}Al measurements should be undertaken in both east- and west flowing catchments. Dating valley-floor sediments and sediment budgeting of Vosgian valleys would bring key elements as well.

Chapitre 3

Post-glacial landscape evolution of the Cleurie catchment (Vosges Mountains, NE France)

Some of the data used in this chapter were produced as part of a Master research project (Fuchs, 2022). We plan to submit this article in the journal **Quaternaire**. Results were presented as a poster at **INQUA2023** (Rome, Italy) and **Quaternaire14** (Rennes, France).

3.1 Introduction

Quaternary glaciations played a major role in the shaping of current fluvial landscapes. The morpho-sedimentary dynamics of fluvial systems evolving in formerly glaciated valleys are conditioned by glacial forms and deposits, which constitute their legacies (Cordier et al., 2017; Mason and Polvi, 2023). To date, studies on post-glacial landscape evolution in areas heavily marked by glaciations have mainly focused on high mountain ranges, such as the European Alps (Bernsteiner et al., 2021; Serra et al., 2022). In many cases, fluvial processes are considered central to the subsequent landscape reshaping, particularly in contexts where paleolakes formed (Tielidze et al., 2021). However, the post-glacial evolution of fluvial systems in formerly glaciated low or mid-mountain ranges has been relatively underexplored (Rixhon, 2021). This is the case of the Vosges in north-eastern France, although glacial imprints on the southern Vosgian landscapes have been already examined since the first half of the 19th century. A local geologist, Henri Charles Hogard, not only immediately embraced Louis Agassiz's (1840) ground-breaking new theory on alpine glaciations but also remarkably transposed it to this low-mountain range (Hogard, 1840, 1842). In his wake, significant

knowledge on past ice dynamics and glacial imprints was added: first at the sub-massif scale (e.g. Collomb, 1847; Grad, 1873; Meyer, 1913; Sittig, 1933), then at the catchment or valley scale mostly (e.g. Dewolf, 1969; Flageollet, 1984; André, 1991; Andréoli et al., 2006).

Against this background, the Upper Moselle catchment (e.g. Nordon, 1931; Flageollet, 1984) and, more specifically, one of its main sub-catchments currently drained by the Cleurie river, has received much scientific attention (Schouler, 1942; Seret, 1967; Salome, 1968; Flageollet et al., 1971; Flageollet, 1988). This catchment is not only located at the heart of the formerly glaciated area but also displays several peculiar geomorphological features at the Vosgian scale. First, it features the highest concentration of glacial and glacial-related deposits among all formerly glaciated catchments: these cover more than 40% of the drainage area (Jautzy et al., 2024b). Second, cores drilled in the valley-floor sedimentary infill of the main stem not only reveal various natures of deposits but also thick sequences that exceed 50 m in the uppermost part of the catchment (BRGM). Third, this small-dimensioned catchment (<80 km²) encompasses the widest range of landforms and/or associated deposits, including glacial (e.g. moraines and erratics), fluvio-glacial (e.g. kame terraces), alluvial (e.g. fluvial terraces), fluvio-lacustrine (e.g. deltas) and marshy (e.g. peat bogs) environments. Despite previous research exploring this unique geomorphological setting (Hogard, 1840; Seret, 1967; Salome, 1968; Flageollet et al., 1971; Flageollet, 1988, 2002), none have studied the post-glacial fluvial evolution of the Cleurie valley or produced numerical datings of the main stages of deglaciation.

Yet, the complex geomorphological evolution of the catchment over the late Quaternary makes it unique at the massif scale to better understand the importance of glacial and post-glacial processes in shaping the Vosges Mountains. However, two key questions are still open:

1. The timing of this evolution is unknown since neither relative age information nor numerical dating is available throughout the catchment. Despite the wealth of glacial landforms and deposits, this lack of chronological data prevents any conclusions about the number of glaciations' imprints preserved in this catchment.
2. The post-glacial fluvial evolution of the hydrographic network, intimately tied to the (de-)glaciation history, is also unknown. This is problematic because the Cleurie catchment, as indicated in Jautzy et al.'s (2024b) dataset covering the whole Vosges Mountains and characterising catchment-wide denudation rates using cosmogenic nuclide measurements from modern riverborne sediments, stands out by the lowest ²⁶Al/¹⁰Be ratio (~ 1.4). This suggests that fluvial sands exported from this specific catchment are in strong cosmogenic unsteadiness, pointing to complex sediment routing systems over long timescales.

Therefore, the Cleurie catchment represents an exceptional study area within the Vosges Massif to characterise and evaluate the impact of glacial legacies on the post-glacial fluvial adjustments of a low-energy mountain stream. The main goals of this study are to (i) evaluate the equilibrium state of the Cleurie's fluvial system and (ii) reconstruct the post-LGM fluvial evolution of the valley. First, building on pioneering geomorphological work conducted from the mid-19th century (Hogard, 1840, 1842) to the late 20th century (Flageollet et al., 1971; Flageollet, 1988), detailed geomorphological mapping of the valley floor was carried out from field surveys and a high-resolution DEM (1 m). Second, the degree of adjustment of the hydrographic network was assessed through the analysis of (i) longitudinal profiles of the main channel and its major tributaries, and (ii) the longitudinal evolution of the bed-surface grain-size of the Cleurie riverbed, with 1400 particles measured (Wolman, 1954). Finally, two key sedimentary profiles in the valley floor were studied, including grain-size measurements and radiocarbon dating of organic matter.

3.2 Study area

3.2.1 The Vosges Mountains

Located in north-easternmost France, the Vosges are a Variscan low-mountain range forming the south-western footwall block of the Upper Rhine Graben (Fig. 3.1A). Neogene uplift is traditionally attributed to the Miocene in relation with compression in the alpine foreland (e.g. Ziegler and Dèzes, 2007) but Harmand and Cordier (2012) also suggested Plio-Quaternary reactivation. Whereas the Vosges Mountains fall upon the moderately tilted Mesozoic formations forming the Paris Basin to the west, their eastern side is clearly bounded by the Upper Rhine Graben's normal fault systems (Fig. 3.1A). The southern and northern parts are made of various Palaeozoic plutonic rocks (mostly) along with secondary metamorphic and sedimentary rocks (crystalline Vosges), and of an homogeneous Triassic sedimentary cover (sandstone Vosges), respectively (Fig. 3.1A).

The Vosges Mountains exhibit a clear topographic gradient evolving from medium elevations to the south (i.e. the Alsatian Grand Ballon peaking at 1424 m above sea level, asl) to much lower mean elevations to the north (~250 m asl, Jautzy et al., 2024b). Moreover, the NNE/SSW-orientation of its main topographic divide (Fig. 3.1A) builds a roughly perpendicular orogenic barrier to the prevailing oceanic western atmospheric flows. Both result in an increase of mean annual rainfall with elevation, i.e. towards the south, and a marked rain shadow effect between the west (wetter) and the east (drier). These massif-wide contrasting features most probably explain the specific spatial pattern of former glaciations and glacially overprinted landscapes shaping the massif's southernmost part, both on its eastern

side (e.g. Sittig, 1933; Andréoli et al., 2006) and western side (e.g. Capot-Rey, 1938; Dewolf, 1969; Flageollet, 1984). Whereas the sandstone Vosges to the north were only locally glaciated (i.e. cirques glaciers, Valenti et al., 2024), the southern part of the crystalline Vosges was repeatedly and prominently glaciated (Fig. 3.1A).

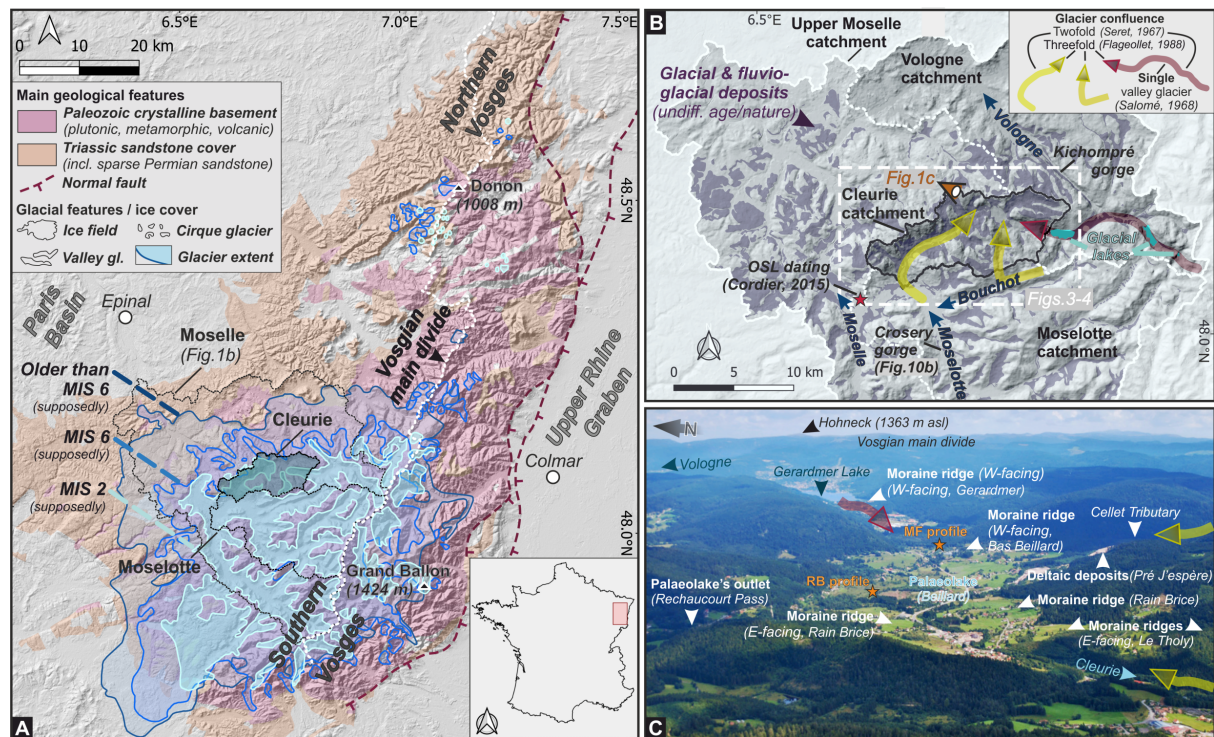


FIGURE 3.1: Geographical setting of the Cleurie catchment. (A) Geological and glacial context of the Southern Vosges. (B) Diverging interpretations concerning number of glaciers that flowed in the Cleurie catchment during LGM. (C) Oblique aerial photograph of the Cleurie Valley, facing towards Gérardmer Lake. Source of geological data: BRGM. Source of DEM: RGEAlt (IGN). Source of glacial extents: Flageollet (2002).

3.2.2 The singular Cleurie catchment

3.2.2.1 Geological and geomorphological background

Located in the south-western part of the massif, the Cleurie valley is a right-bank tributary of the Moselotte river, thereby representing one of the main sub-tributaries of the Upper Moselle amidst the crystalline Vosges (Fig. 3.1B). Palaeozoic plutonic rocks, i.e. granites and leucogranites, mostly occur with subordinate occurrences of metamorphic (migmatites) and sedimentary rocks (sandstones and conglomerates). The Cleurie's NNE/WSW-elongated drainage area amounts to 78 km² at its outlet. Catchment's mean elevation, slope, normalised steepness index and annual precipitation amount to ~700 m, 12°, 42 m^{0.9} and 1360 mm, respectively (Jautzy et al., 2024b). From the source (~810 m) located directly westward of the Gérardmer Lake to its confluence with the Moselotte in Saint-Amé (~390 m), the main

stem's length and mean longitudinal slope amounts to 22.5 km and 2.0 %, respectively. The main valley's morphology is overall deeply incised in the crystalline rock formations (Fig. 3.1C), with maximum height differences between valley-floor and interfluvies reaching ~ 350 m in the lowermost reach. However, the valley-floor's morphology strongly varies despite the rather limited main stem's length, ranging from large and flat areas whose width roughly ranges between ~ 500 and 1000 m (Fig. 3.1C), to narrow and steep reaches.

3.2.2.2 Unordinary glaciated landscape and complex glacial legacy

Along with other features detailed hereinafter, this last observation bears witness of abundant and well-preserved glacial imprints within the Cleurie catchment. First, harmonised geological mapping of the French Geological Survey (BRGM) at the massif scale reveals the highest proportion of sediments within this catchment, regardless of their nature (i.e. covering almost half of the complete draining area). Even more significant is the predominant proportion, among these, of undifferentiated glacial and fluvio-glacial sediments: $>42\%$ (Fig. 3.1B; Jautzy et al., 2024b). Second, the main valley-floor sedimentary infill highlights thick and variable sequences locally exceeding 50 m in the catchment's uppermost part (BRGM's BSS database; <https://infoterre.brgm.fr/page/banque-sol-bss>). Third, despite its rather small area, the Cleurie catchment most probably includes the highest concentration and widest array of glacial and glacial-related landforms and/or deposits at the massif scale. They encompass (i) till material, (ii) end moraines (Fig. 3.1C), (iii) erratic boulders, (iv) kame terraces, (v) pro-glacial deltaic/glacio-lacustrine deposits (Fig. 3.1C) and (vi) peat bogs (Fig. 3.2).

In the wake of Hogard's (1840; 1842) precursory works containing, among others, valuable massif-scale observations on erratics and moraines along with scientific sketches and descriptions (Fig. 3.2), the uniqueness of the Cleurie is reflected by the unequalled number of geomorphological studies performed in this catchment at the Vosgian scale (Meyer, 1913; Nordon, 1931; Capot-Rey, 1938; Schouler, 1942; Seret, 1967; Salome, 1968; Flageollet et al., 1971; Flageollet, 1988). The main conundrum concerned former ice dynamics, supposedly during the last glacial cycle, and questions related to the number of valley glaciers and the nature of deposits occurring in the Cleurie catchment (Fig. 3.1B). On the one hand, Salome (1968) assumed a single E-W valley glacier originating from the Vosgian main divide in the uppermost Vologne catchment (Fig. 3.1B). The vast majority of landforms and correlated deposits along the main valley were interpreted as kame terraces/ridges due to the key role presumably attributed to meltwater channels and fluvio-glacial processes (Fig. 3.3A). On the other hand, Seret (1967) assumed a valley-glacier convergence in the upper Cleurie catchment. In addition to the westward-moving glacier, a second eastward-moving glacier originating from the Moselotte valley, i.e. flowing upslope the Cleurie valley, was postulated (Fig. 3.1B). The same

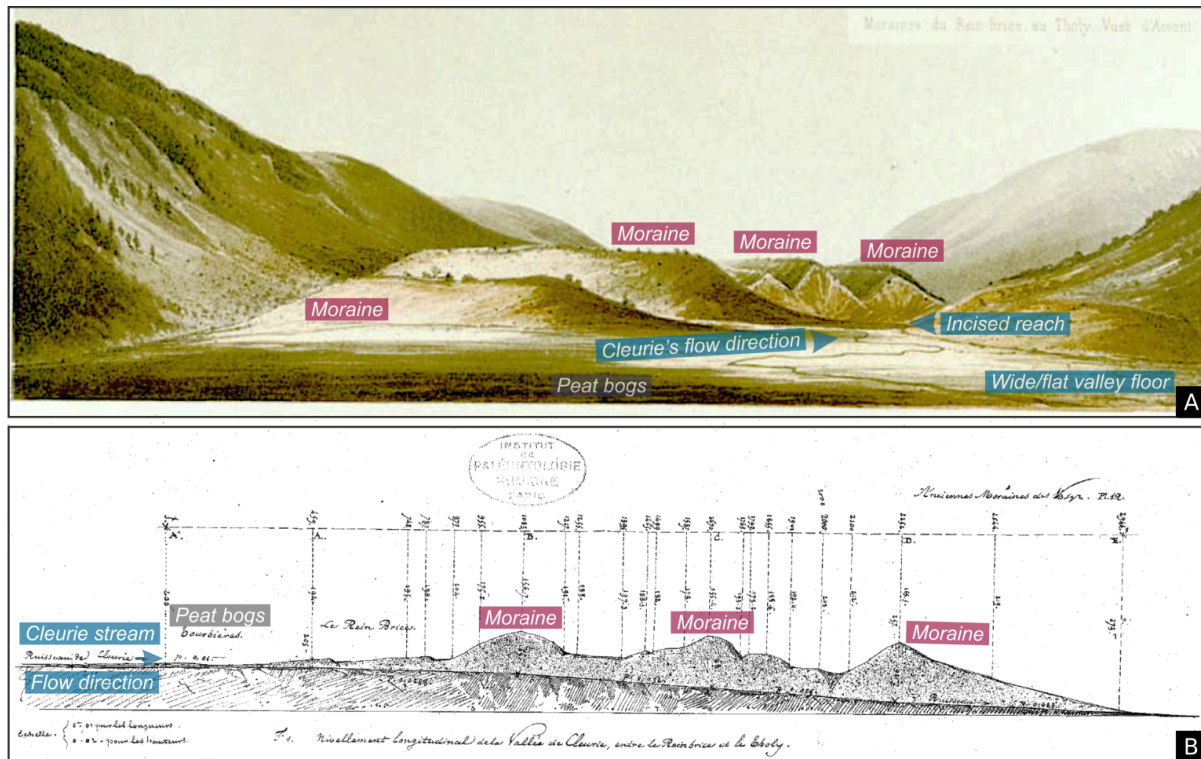


FIGURE 3.2: First scientific illustrations of Le Tholy's morainic complex. (A) Artistic watercolor depicting the sequence of presumed frontal moraines of Le Tholy, looking downstream in the valley. Adapted from Hogard (1840). (B) Cross-sectional profiles depicting the sequence of presumed frontal moraines of Tholy. Adapted from Hogard (1842).

landforms/deposits were interpreted as eight distinct end-moraine ridges due to the key role presumably attributed to glacial processes (Fig. 3.3B). This hypothesis was later refined by Flageollet (1988), who assumed a threefold glacier convergence, with the third valley glacier bypassing the Cleurie's southern interfluve *via* the Cellet's tributary valley (Figs. 3.1B-C, 3.3C). This hypothesis of a threefold glacier convergence relies on morphological observations, i.e. west- vs east-facing end-moraine ridges along the main valley (Figs. 3.1C, 3.3C), and sedimentological data, i.e. clast lithological analysis revealing different proportions of crystalline rocks, hence different origins of glacial material (Seret, 1967; Flageollet et al., 1971). Note that these valley glaciers, while converging into the upper Cleurie catchment, seemingly did not merge into a single glacier. As glacio-lacustrine deposits occur within three distinct altitudinal ranges throughout the catchment (i.e. ~ 675 , 630 and 550 m asl), three generations of palaeolakes could be accordingly inferred (Flageollet et al., 1971). Maximal E-W extension of the highest palaeolake – the so-called Beillard palaeolake – reached 6 km and its outlet was located on the Cleurie's northern interfluve, the so-called 'Rechaucourt pass', flowing to the Vologne catchment (Fig. 3.1B; Flageollet, 1988).

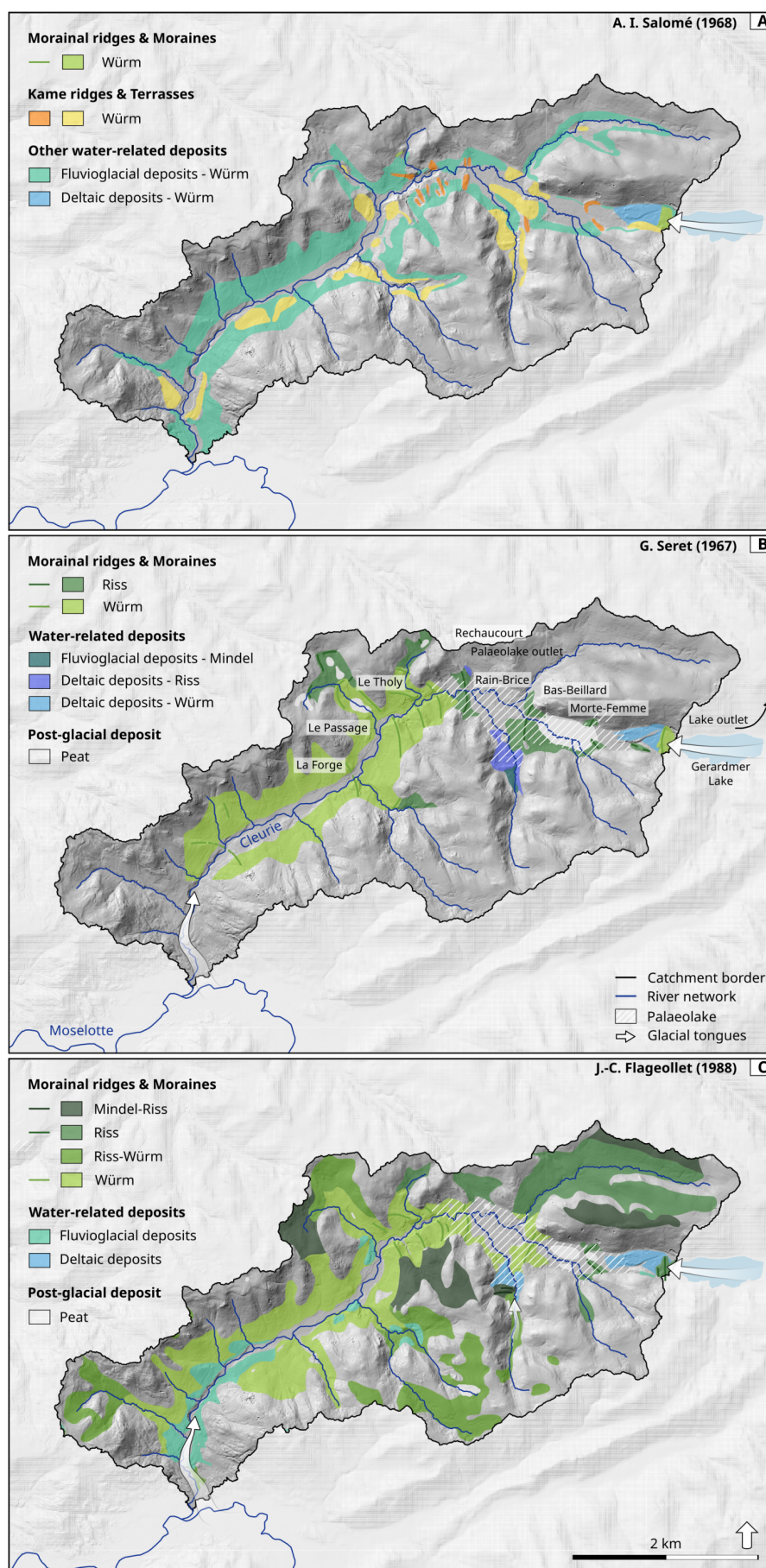


FIGURE 3.3: Geomorphological maps of the glacial-related deposits in the Cleurie catchment, according to (A) Salomé (1968), (B) Seret (1967) and (C) BRGM geological map (Flageollet, 1988).

3.2.2.3 Two pending key questions

However, two key aspects were neglected. First, numerical age estimates on Cleurie's deposits completely lack, thereby impeding a sound temporal understanding of the catchment's Quaternary evolution. This is all the more astonishing given their spatial clustering and their wide diversity in nature, i.e. glacial, fluvio-glacial, glacio-lacustrine and fluvial deposits. If (i) the Vosges Massif's south-western part (including our study area) is supposed to have been glacially shaped during (at least) three distinct cold periods (Fig. 3.1A) and (ii) several phases of glacier retreat were suggested in the Cleurie catchment during the supposedly last glacial cycle (Flageollet, 1988), all chronological outcomes and reconstructions unfortunately remain mere assumptions.

Second, observations on fluvial features within the catchment are almost completely absent. Meyer (1913), Nordon (1931), Capot-Rey (1938), Schouler (1942), Seret (1967), Salome (1968), and Flageollet (1988) extensively focused on glacial(-related) deposits but almost completely disregarded the post-glacial fluvial overprint on the Cleurie's landscape. Only Flageollet et al.'s (1971) comprehensive study, though focusing on glacial deposits at the valley scale, pointed out several fluvial terrace remnants along its lowermost reach. This fluvial overprint not only involves fluvial deposits which blanket the valley-floor but also the post-glacial evolution of the hydrographic network which should have been strongly impacted by this probable threefold glacier convergence. In this respect, Jautzy et al.'s (2024b) dataset quantifying catchment-wide denudation rates using cosmogenic nuclide measurements from modern riverborne sediments at the massif scale is particularly interesting. The Cleurie exhibits, by far, the lowest $^{26}\text{Al}/^{10}\text{Be}$ ratio (~ 1.4), meaning that modern fluvial sands exported from this specific catchment are in strong cosmogenic unsteadiness on 10^5 -years timescales. This observation not only points to complex sediment routing systems within the catchment but is also well in line with a singular, at the Vosgian scale, glacial legacy.

3.3 Methodology

3.3.1 Catchment-scale geomorphological analysis

3.3.1.1 Geomorphological mapping

This study aims to produce an updated geomorphological map of the Cleurie catchment by incorporating two key elements: (1) previously unmapped fluvial morpho-sedimentary units covering the valley-floor and (2) previously mapped glacial-related units located at higher elevations along the hillslopes of the main stem and some tributaries. Regarding the latter,

we relied on Flageollet et al.'s (1971) map, as it is the only one that identifies the nature of deposits, unlike other available maps (e.g. Flageollet, 1988). According to Flageollet et al. (1971), these deposits may be related with the gradual lowering of the former paleolake(s). Because it likely influenced subsequent fluvial adjustments, we argue that integrating these deposits into our updated geomorphological map could enhance our understanding of the post-LGM relative chronology in the Cleurie catchment.

We used a LiDAR-derived digital elevation model (DEM) with a 1-meter resolution provided by the French National Institute of Geography (IGN). Using QGIS, we generated a slope raster to identify flat surfaces. Fluvial morpho-sedimentary units were delineated based on their relative elevation compared to the current thalweg and their levelness, distinguishing between modern floodplains and low-lying fluvial terraces. Field surveys were conducted to confirm the nature of identified flat surfaces on the valley-floor.

Regarding previously mapped glacial-related units (Flageollet et al., 1971), we reassessed them using a LiDAR-derived DEM. Georeferencing of Flageollet et al.'s (1971) map was achieved using an orthorectified ancient (mid-20th) topographical map (SCAN50, IGN) through a second polynomial transformation using 10 Ground Control Points (GCPs), resulting in a RMSE of 6.99 m. The deposits were initially digitised according to Flageollet et al.'s (1971) mapping. Given the nature of the deposits identified by the authors (i.e. essentially glacio-lacustrine) and their resulting characteristics (i.e. low surficial slope, between 10 and 20°), their contours were refined using the shaded DEM and a derived slope raster. Some deposits were reclassified into different types, such as morainal ridges. Additionally, due to quarrying since the publication of Flageollet et al.'s (1971) map, the former contours of some deposits were extrapolated to account for changes.

3.3.1.2 Longitudinal profiles and bed-surface grain-size variations

The Cleurie catchment has been significantly affected by glaciations, which have shaped the landscapes of the valley and its tributaries, depositing large quantities of glacial-related sediments. It has been demonstrated that the Cleurie catchment area is in strong cosmogenic disequilibrium (Jautzy et al., 2024b). Additionally, the existence of palaeolake(s) may have locally and temporarily altered the base level in the valley. It is thus assumed that a sedimentary imbalance may be visible in the longitudinal profile of the Cleurie and its tributaries. To analyse the current degree of adjustment of the Cleurie river and its tributaries, we extracted and analysed their longitudinal profiles. We applied the 'lsdtt-chi-mapping' function of the LSDTopoTools software (Clubb et al., 2014) to a 5 m-resolution DEM provided by the IGN to automatically extract the hydrographic network of the Cleurie catchment. Resolution of the DEM was chosen to achieve maximum accuracy with minimum topographical noise

(Persendt and Gomez, 2016). The extracted network is a point shapefile (point=pixel) in which each point contains the following attributes: distance to confluence, elevation, drainage area. We set the minimum threshold of drainage area required to generate a stream at 0.25 km^2 and we only retained those streams with a minimum length of 1.5 km. As a result, a total of 16 (sub)tributaries were extracted (#2-17; Fig. 3.4) in addition to the Cleurie's main stem (#1; Fig. 3.4). Longitudinal local slope of the whole river network was then computed at each point, i.e. each 5 m, using a fixed horizontal distance of 100 m. Logarithmic plots of longitudinal river gradient or slope versus drainage area, i.e. slope-area data, are useful to detect spatial variability in steepness index within river networks (Wobus et al., 2006). At steady state, the latter is influenced by e.g. rock uplift rate or erodibility, climatic variables, channel and/or sediment properties (Whipple, 2004). We thus produced slope-area data along the main stem. As tributaries input in drainage area typically induces gaps along the x-axis (Mudd et al., 2016), confluences were represented to facilitate the combined lecture between these data and the long profile. We followed Wobus et al.'s (2006) categorisation of river long profiles: adjusted (i.e. concave), linear, knickpoint and hanging.

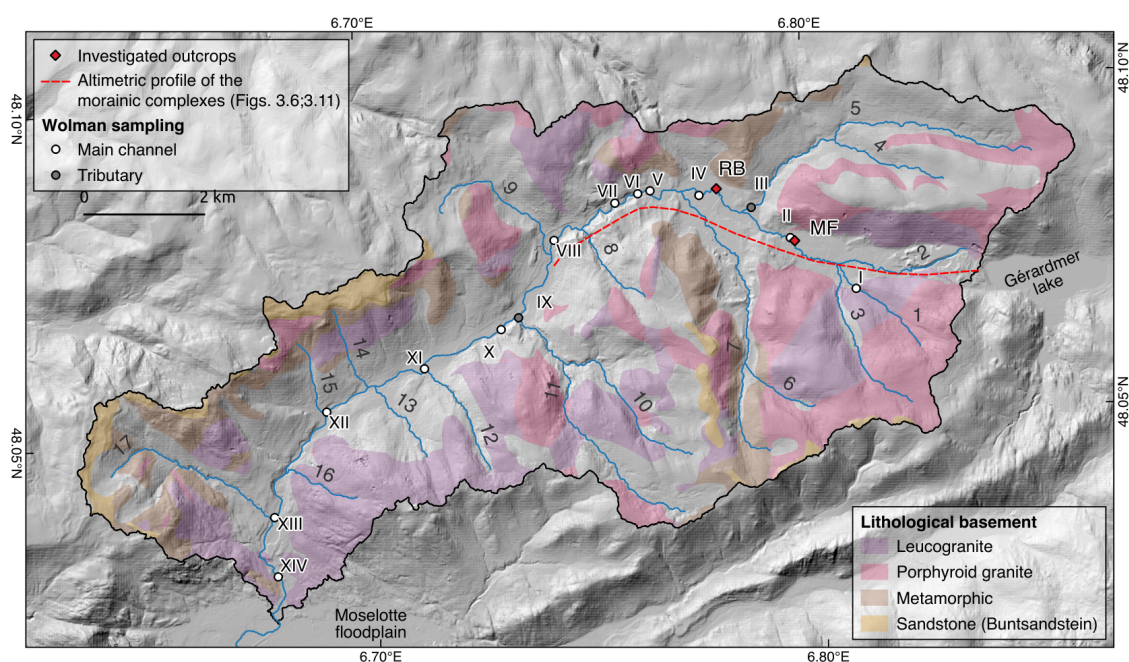


FIGURE 3.4: Hydrological network of the Cleurie catchment and location of the investigated outcrops and the Wolman sampling. Tributaries are numbered from upstream to downstream. The Cleurie's main channel is numbered #1. Lithological basement is based on the BRGM geological map. Areas not covered correspond to Quaternary deposits.

Along-stream grain-size variations within the main channel were assessed *via* twelve *in situ* bedload measurements (samples #I,II,IV-VIII,X-XIV; Fig. 3.4). As the Wolman's (1954) pebble-count method suits the study of gravel-bed rivers, such as the Cleurie, we randomly selected 100 clasts at each site and measured their b-axis but four points regarding the sampling

approach should be noted. First, we avoided river reaches strongly impacted by man-induced modifications. Second, along-stream distance between each site commonly ranges between 1.5 and 2 km to capture the variability along the whole main stem. However, we densified our measurements where the Cleurie valley-floor bears witness of significant glacial legacy, i.e. where the river cuts through the main deposit complexes. Third, boulders (>256 mm) are abundantly strewn across the modern riverbed, precisely because of this same legacy. We opted to integrate these coarsest clasts in our sampling approach. Despite (i) measurement of their apparent b-axis could only be performed, resulting in a minimum assessment of their size, and (ii) their impact on the spatial variability of the bed-surface grain-size, hence the representativeness of the Wolman pebble-count method (Bunte and Abt, 2001), they allowed documenting the presence and relative abundance of glacial(-related) deposits in the riverbed. This is well in line with our aim to understand the relationship between fluvial and glacial(-related) landforms. Four, we also sampled two main tributaries directly upstream of their confluence (i.e. tributaries #5 and 11, samples #III,IX; Fig. 3.4) to check their input material for potential main stem's bedload coarsening or fining (e.g. Brierley and Fryirs, 2013).

3.3.2 Sedimentological and chronological analysis (section scale)

3.3.2.1 Fieldwork: logging and grain-size/¹⁴C sampling

Two sedimentary profiles were thoroughly documented along the Cleurie valley-floor (Fig. 3.4): the so-called Morte-Femme (MF: 48°4'35.2963"N, 6°47'48.7554"E) and Rain-Brice sequences (RB: 48°5'5.1434"N, 6°46'47.6117"E). MF and RB are natural outcrops in the outer, retreating banks of current meanders, located in the upper and lower reach of the area formerly filled by the paleolake (Fig. 3.3C; Fig. 3.4). We first performed a full characterisation, sedimentological description and 1/20-scale logging of both profiles. Coordinates and elevation of the surface of each profile were measured using a Trimble GeoXH 6000 GNSS combined with a Zephyr 2 antenna, allowing a centimetric accuracy. Grain-size sampling was then performed with a vertical resolution corresponding to the stratigraphic units previously identified. Thick units (>10 cm) or units presenting some variability were sampled at multiple locations. Collection of samples for radiocarbon dating was performed according to the prevailing grain-size along both profiles. Macroscopic plant or woody remains embedded in sandy units, visible to the naked eye after clearing the sections, were present in RB (i.e. an *in situ* tree trunk) and the upper part of MF. These were directly collected in plastic bags. For the lower part of MF consisting of clayey units, no macroscopic material was directly visible. Five bulk parallelepipeds (~10 cm-high, 10 cm-wide and 20 cm-long) were carefully extracted after clearing the section for later sieving in the laboratory.

3.3.2.2 Labwork: grain-size measurements and ^{14}C dating

Grain-size analysis was performed in the Laboratoire d'Analyse des Sols at Strasbourg University. Samples were oven dried for 8 days at 40°C and sieved at 2 mm. The grain-size distribution was measured by laser granulometry (Beckman Coulter LS230), either with the Aqueous Liquid Module for silty and clayey samples, or with the Tornado Module for sandy samples. The clayey bulk parallelepipeds collected at the lower part of the MF section were cleaned for edge contamination, then vertically separated into two parts. One part was subsampled with a 5 cm vertical resolution. Subsamples were dissolved in distilled water and water-sieved at 250 μm to collect organic material. All collected elements were observed under the binocular and sent to the CEDAD laboratory (Università del Salento, Italy) for AMS radiocarbon dating. Age estimates were calibrated using the software OxCal v.4.4 based on the last atmospheric dataset (Reimer et al., 2020).

3.4 Results

3.4.1 Geomorphological map

3.4.1.1 Integration of valley-floor fluvial features and valley segmentation

First, the Cleurie's valley-floor morphology globally features wide and flat reaches (unconfined valley) alternating with narrower and steeper reaches (confined valley; Fig. 3.5). Second, two successive levels of widespread (i.e. present in several reaches; see hereinafter) fluvial units can be distinguished according to their mean relative elevations (with 1σ) above the modern thalweg (from the lowest to the highest): 1.5 (± 0.3) and 3.3 (± 1.1) m. They match the modern river floodplain and a lower terrace level. Third, the river pattern also ranges from a predominant straight channel to a predominant meandering channel, both being either coupled to or decoupled from hillslopes. Based on these threefold spatial variations, i.e. narrow vs wide valley-floor, occurrence (or not) of the lower terrace level, coupled/decoupled straight vs meandering channel, the main valley can be subdivided into five roughly homogeneous reaches (≥ 2 km long each) distributed up-downstream as follows (Fig. 3.5):

- ~ 2 km long first reach (from the source downstream to roughly 0.6 km upstream of the confluence with tributary #3): narrow valley-floor, absence of floodplain or terraces, straight channel directly coupled to hillslopes (i.e. headwaters);
- ~ 4 km long second reach (downstream to the confluence with tributary #7): up to 1 km wide valley-floor, continuous and widespread occurrence of the lower terrace,

predominant meandering channel (short wavelength) fully decoupled from hillslopes, excepted along a ~ 0.4 km long intermediate reach through the Bas-Beillard's complex;

- ~ 3.5 km long third reach (downstream to the confluence with tributary #9): narrow valley-floor, occurrence of the lower terrace, laterally confined, predominant meandering channel (long wavelength) partly coupled to hillslopes;
- ~ 4.5 km long fourth reach (downstream to the confluence with tributary #14): up to 0.5 km wide valley-floor, almost total absence of the lower terrace, predominant straight channel almost fully decoupled from hillslopes;
- ~ 4.5 km long fifth reach (downstream to the Moselotte floodplain): relatively narrow valley-floor (maximal width reaching 0.2-0.25 km), occurrence of the lower terrace and absence of the modern floodplain, predominant straight channel partly coupled to hillslopes.

3.4.1.2 Update of previously mapped glacial-related landforms

We have included the 'glacial-related' deposits mapped by Flageollet et al. (1971). According to the authors, they are all characterised by a remarkably flat surface and lie at distinct altitude levels. The localisation of these deposits have been checked and updated from the DEM, mainly on the basis of the flatness criteria highlighted by the authors (i.e. slope between 10 and 20°). Two types of deposits are distinguished in their work: glacio-lacustrine (deltaic) deposits and fluvio-glacial terraces (Fig. 3.5). These are distributed in the Cleurie catchment as follows. The glacio-lacustrine deposits are divided into four distinct altitude levels. Those from the lower level (533 ± 16 m) are located exclusively in reach 4, downstream of the Tholy's morainic complex, mainly at footslopes. Those from the intermediate level (616 ± 12 m) are located in reaches 2 and 3 and upstream of reach 4. Those from the upper level (664 ± 8 m) are located in reaches 1, 2, 3 and upstream of reach 4 (Fig. 3.5A). The 'Bas-Beillard' level (598 ± 3 m) is located in reaches 2 and 3 (Fig. 3.5B). As for the fluvio-glacial terraces, the 'Lower Cleurie' terrace (457 ± 20 m) is located in reach 5 (Fig. 3.5A), mainly on the right bank of the Cleurie.

3.4.2 Long profiles and bed-surface grain-size

3.4.2.1 River network

The main stem's long profile has a mean slope of 2.0% (standard deviation of 2.8%) but clearly display several irregularities (Fig. 3.6). Four shifts in linear arrays of slope-area data,

Glacial-related units, modified from Flageollet and Hameurt (1971)
(absolute mean elevation asl, $\pm 1\sigma$)

Glacio-lacustrine

- Lower level
(533 $\pm$ 16 m)
- Intermediate level
(616 $\pm$ 12 m)
- Upper level
(664 $\pm$ 8 m)
- Bas-Beillard level
(598 $\pm$ 3 m)

Fluvio-glacial

- Lower Cleurie terrace
(457 $\pm$ 20 m)

Valley-floor fluvial units
(relative elevation
above thalweg, $\pm 1\sigma$)

Alluvial

- Modern floodplain
(1.5 $\pm$ 0.3 m)
- Lower terrace
(3.3 $\pm$ 1.1 m)

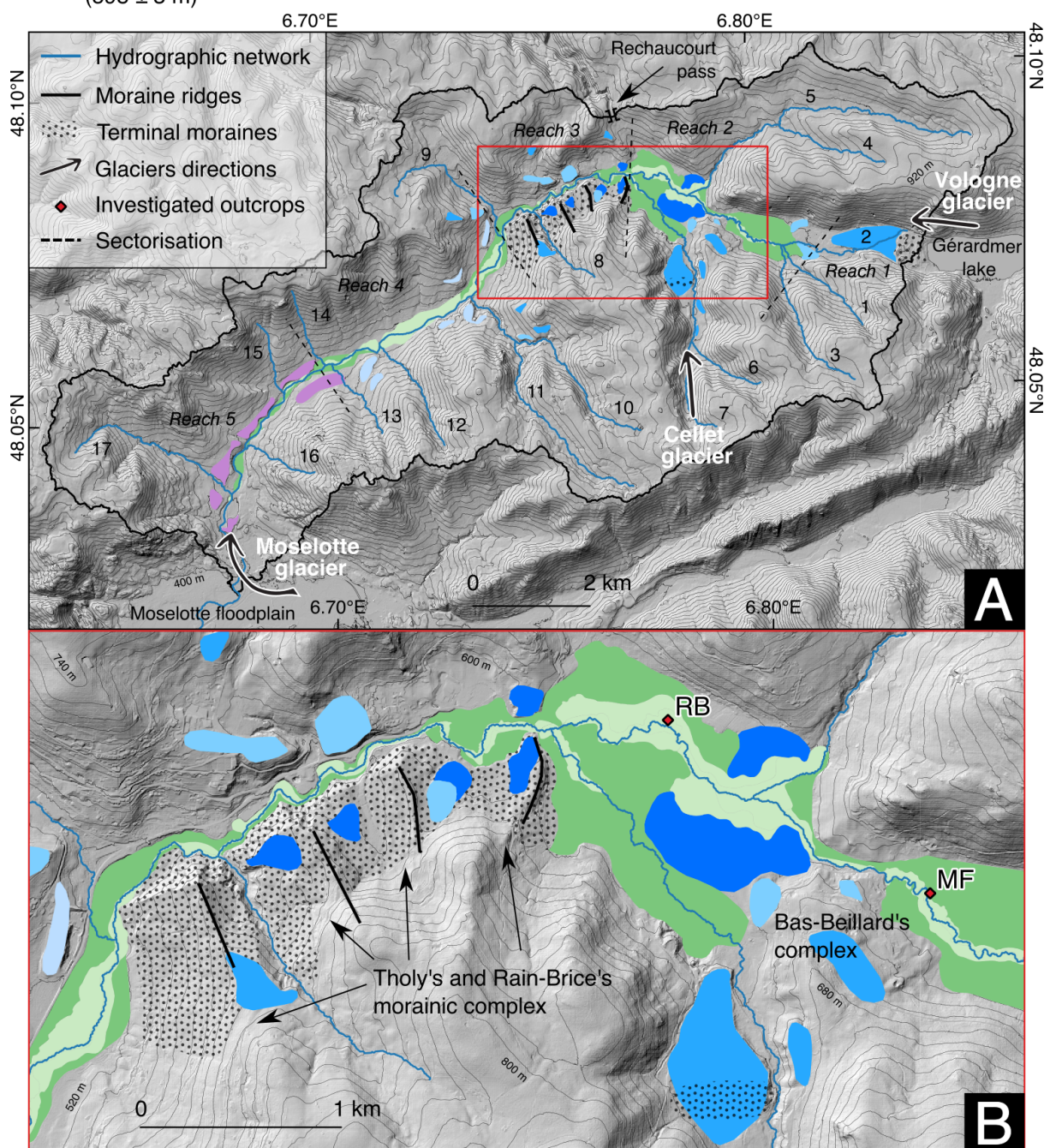


FIGURE 3.5: Updated geomorphological map of the Cleurie catchment with reappraisal of previously mapped glacial-related (Flageollet et al., 1971) and fluvial morpho-sedimentary units. Isolines are separated by 20 meters vertically.

highlighted by local slope increases as a function of drainage area (Fig. 3.6A), coincide with two knickpoints and two knickzones conspicuous along the Cleurie's long profile (Fig. 3.6B). According to the river-course segmentation above, the two localised knickpoints (<50 m in length) occur in the first (uppermost) and fifth (lowermost) reaches. The two ~0.5 and 2 km long knickzones occur in the second and third reaches, respectively. Altogether, nine left-bank (#3, 6, 7, 8, 10, 11, 12, 13, 16) and seven right-bank (sub-)tributaries (#2, 4, 5, 9, 13, 14, 15, 17) form – alongside the main stem – the Cleurie's current river network. Morphometric variability is attested in this catchment as all categories of river long profiles are represented (Wobus et al., 2006): three (sub-)tributaries are adjusted (#13, 14, 17), four are straight (#6, 10, 15, 16), seven have slope breaks (i.e. knickpoints, #2, 3, 4, 7, 8, 9, 12) and two are hanging valleys (#5, 11).

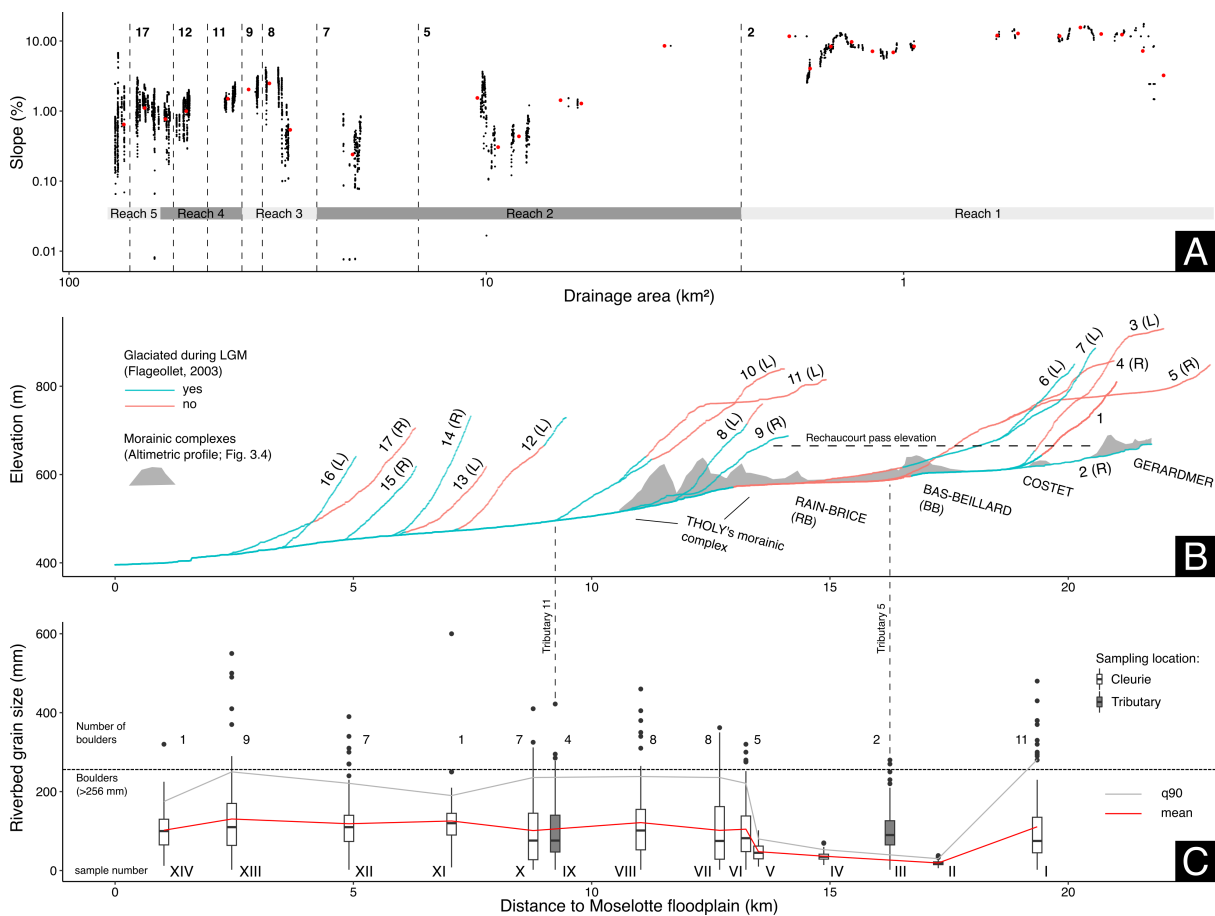


FIGURE 3.6: Longitudinal morphometry of the Cleurie river. (A) Slope/area plot extracted from Cleurie's longitudinal profile. Red points correspond to a moving average over 250 m. The vertical dashed lines localise tributaries' confluences, identified by their respective numbers. (B) Long profiles of the Cleurie river and main tributaries. Moraines profile was extracted from the transect represented in Figure 3.4. (C) Longitudinal evolution of the bed-surface grain-size of the Cleurie river measured with Wolman pebble-count.

3.4.2.2 Along-stream variations of bed-surface grain-size

Twelve assessments of bed-surface grain-size in the main stem (i.e. b-axis length of 1200 *in situ*-measured pebbles) yield a mean (with 1σ) and median (i.e. D_{50}) grain-size amounting to 93 (± 82) mm and 70 mm, respectively (Fig. 3.6C). D_{90} (200 mm) is roughly three times higher than D_{50} whilst boulders (>256 mm) represent $\sim 0.5\%$ of all sampled particles and the apparent b-axis of the largest clast measures ~ 600 mm. Importantly, order-of-magnitude along-stream variations in D_{50} and D_{90} are observed; they range from 19 to 120 mm and 30 to 250 mm, respectively. On the one hand, three sites in a row located upstream along the main stem, i.e. sites II, IV and V, consistently show low D_{50} and D_{90} (<50 and ≤ 80 mm, respectively) along with the absence of clasts >256 mm. On the other hand, all measurements performed downstream (VI- $\rightarrow$ XIV) consistently show high D_{50} and D_{90} (≥ 75 mm and ≥ 190 mm, respectively) along with the presence of at least one clast >256 mm at each site, with a peak proportion of boulders of 9% in the lowermost site (Fig. 3.6C). Note also the sharp grain-size transition between sites V and VI as they are located only 0.2 km apart from each other. Mean grain-size, D_{50} and D_{90} of each tributary (#5 and 11; sampling sites III and IX) are in the same range than those of the main stem. In tributaries #5 and 11, they amount to 103 (± 59), 90 mm and 200 mm, and 97 (± 74), 76 and 180 mm, respectively. Boulders represent $\sim 0.3\%$ of all sampled particles there.

3.4.3 Stratigraphical, grain-size and chronological results

3.4.3.1 MF profile (Morte-Femme)

Owing to active lateral erosion in the meander's outer bank, the MF profile reveals an up to 2 m high sequence continuously observable along ~ 20 m (Fig. 3.7). Absolute elevations of the base and the top of the profile are 605.3 and 607.2 m, respectively. Eight stratigraphic units are identified along the 1.92 m high logged profile; they are described from bottom to top:

- Unit A (1.92 to 1.25 m; base not observed): lacustrine rythmites. Alternation of plastic, light-pink clayey laminae and light-green silty laminae;
- Unit B (1.25 to 0.30 m): fluvial deposit. Matrix-supported layers, containing various proportions of clasts;
 - Subfacies Ba (1.25 to 1.20 m): clast-supported bed, made of heterometric inclusions ranging from centimetric granules to decimetric pebbles;
 - Subfacies Bb (1.20 to 1.10 m): matrix-supported beige-coloured layer, characterised by a fining-upward sequence, ranging from granules to medium sand, sharp boundary to underlying layer;

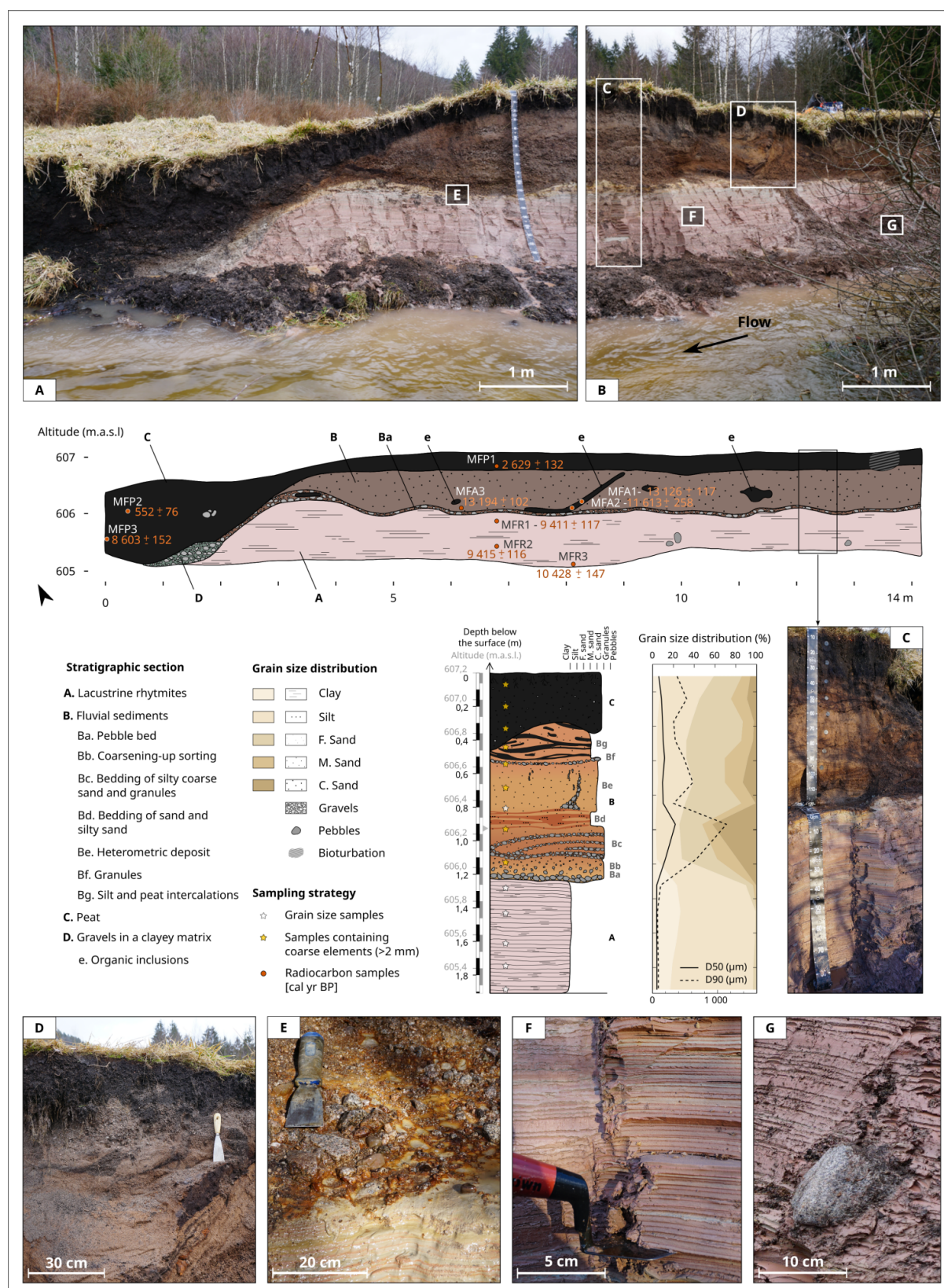
- Subfacies Bc (1.10 to 0.95 m): reddish brown beds of fine granules and coarse sand;
- Subfacies Bd (0.95 to 0.80 m): alternation (i.e. cross-bedding) of black silty sand laminae and light brown/beige medium sand laminae, sometimes slightly silty;
- Subfacies Be (0.80 to 0.55 m): light brown/beige silty sand layer, containing organic inclusions (roots) and vertical lenses of centimetric granules;
- Subfacies Bf (0.55 to 0.50 m): clast-supported bed, made of heterometric material ranging from centimetric granules to coarse sand;
- Subfacies Bg (0.50 to 0.30 m): cross-bedded sandy layers, ranging from ochre/beige fine sand to black organic-rich silty sand;
- Unit C (0.30 to 0.00 m): black organic-rich peat layer of various thickness, with silty matrix frequently including roots and scattered granules.

The profile is tripartite according to the grain-size results performed in the matrix-dominated layers. On the one hand, the uniform and well-sorted unit A from 1.90 to 1.25 m (sampled five times) is predominantly made of silt (75%) and secondarily of clay (22%), sand being almost absent (3%). D_{50} and D_{90} remain almost invariant around 6.3 and 35.4 μm , respectively. On the other hand, all units from 1.25 m to the sub-surface are predominantly made of sand (eight samples); its proportion peaks in subfacies Bb-Bc ($\sim 73\%$) and gradually decreases towards the top (the uppermost sample showing 41% of sand). Whilst particles coarser than 2 mm are observed in all nine samples but one, the rest is mostly composed of silt (22 to 50%) and clay to a much lesser extent ($\sim 5\%$). Coarse sand increases in unit C and reaches its peak proportion at its top ($\sim 9\%$) materialised by the highest D_{50} and D_{90} of the whole profile (83 and 494 μm , respectively).

Radiocarbon dating results give an age of ca. 10'430 to 9'410 cal. BP for the light-pink clayey laminae and light-green silty laminae of sedimentary unit A, which are interpreted as lacustrine rhythmites, with three radiocarbon dates in stratigraphic order. Radiocarbon dates coming from the overlying pebbly-sandy unit B, interpreted as fluvial deposits, are quite older (ca. 13'190 to 11'610 cal. BP) and show a stratigraphical inversion between the middle sample and the lower and upper ones. Organic-rich unit C displays three inconsistent radiocarbon dates, at ca. 8'600, 550, and 2'630 cal. BP from the lower to the upper sample (Fig. 3.7, Table 3.1).

3.4.3.2 RB profile (Rain-Brice)

Owing to active lateral erosion in the meander's outer bank, the RB profile reveals an up to 1.2 m high sequence continuously observable along ~ 20 m (Fig. 3.8). Absolute elevations of the base and the top of the profile are 583.0 and 584.2 m, respectively.



Sample ID	Sedimentary unit (Fig. 3.7)	Lab. Code (CEDAD)	Depth [cm]	Radiocarbon age [^{14}C yr BP]	Calibrated age range [cal yr BP] (95.4 %)	Material
Morte Femme						
MFP1	C	LTL31147	25	2573 ± 40	2760-2497	charcoal
MFP2	C	LTL22247	45	494 ± 45	628-476	wood
MFP3	C	LTL22248	82	7815 ± 45	8754-8451	bark
MFA1	Be	LTL22250	85	11216 ± 55	13243-13009	wood
MFA2	Be	LTL22249	90	10091 ± 50	11870-11355	wood
MFA3	Be	LTL31149	110	11264 ± 55	13295-13092	wood (pinus)
MFR1	A	LTL31152	130	8398 ± 50	9528-9294	fibrous organic material (grass fragments)
MFR2	A	LTL31153	165	8410 ± 50	9531-9299	fibrous organic material (grass fragments)
MFR3	A	LTL22246	220	9267 ± 45	10575-10281	charcoal & organic fragments
Rain-Brice						
RB	-	LTL22251	60	1756 ± 45	1730-1543	in-situ tree trunk

TABLE 3.1: Radiocarbon dating results. Sample ID, respective sedimentary units for the MF section (A = lacustrine rythmites, Be = heterometric deposit in fluvial sediments, C = peat ; Fig. 3.7), Lab. Code, Depth, uncalibrated age and respective error (1σ), age range, calculated with OxCal 4.4 (Bronk Ramsey, 2009), using the IntCal20 calibration curve (Reimer et al., 2020), and material dated using ^{14}C .

The 1.2 m high Rain-Brice (RB) profile corresponds to a ~ 20 m long natural outcrop located on the right bank of a meander and resulting from the Cleurie's channel lateral erosion. Two main elements stand out from the outcrop: a curved structure and a tree trunk occluded in the bank (Fig. 3.8). The curved structure corresponds to a bent form of fine dark layers, crosscutting the sediments below (Fig. 3.8). Absolute elevations of the base and the top of the profile are 583.0 m and 584.2 m, respectively. RB is a fully alluvial deposit, showing a general fining-up sorting from pebble inclusions at the bottom to sandy and silty layers at the top. We identified 3 stratigraphic units (Fig. 3.8):

- Unit A (1.20 to 1.10 m) is a heterometric layer, from coarse sand to pebbles of 3-4 cm, slightly rounded-edged.
- Unit B (1.10 to 0.20 m) is a sandy silt layer, in which 6 subfacies are identified:
 - Subfacies Ba (1.10 to 1.00 m): medium to coarse sand, well sorted. Poorly cohesive, this layer is beige coloured and contains several organic remains;
 - Subfacies Bb (1.00 to 0.95 m): homometric sandy silt, organic and black-coloured. The layer contains some organic remains;
 - Subfacies Bc (0.95 to 0.85 m): well sorted fine silty sand, pale beige and presenting black horizontal streaks;
 - Subfacies Bd (0.85 to 0.55 m): coarsening-up sorting, ranging from silt to coarse sand. The whole unit is greyish brown coloured. This stratigraphic unit contains a well-preserved tree trunk, laying in a horizontal position;

- Subfacies Be (0.55 to 0.50 m): heterometric layer of medium to coarse sand, beige coloured;
- Subfacies Bf (0.50 to 0.20 m): thick layer of sandy silt, well sorted and pale brown/beige coloured. Rare roots are scattered;
- Unit C (0.20 to 0.00 m): silty sandy backfill, grey/brown coloured. Some slightly rounded-edged cobbles are scattered, as well as roots. An irregular iron oxidised edging is visible at the bottom of the layer.

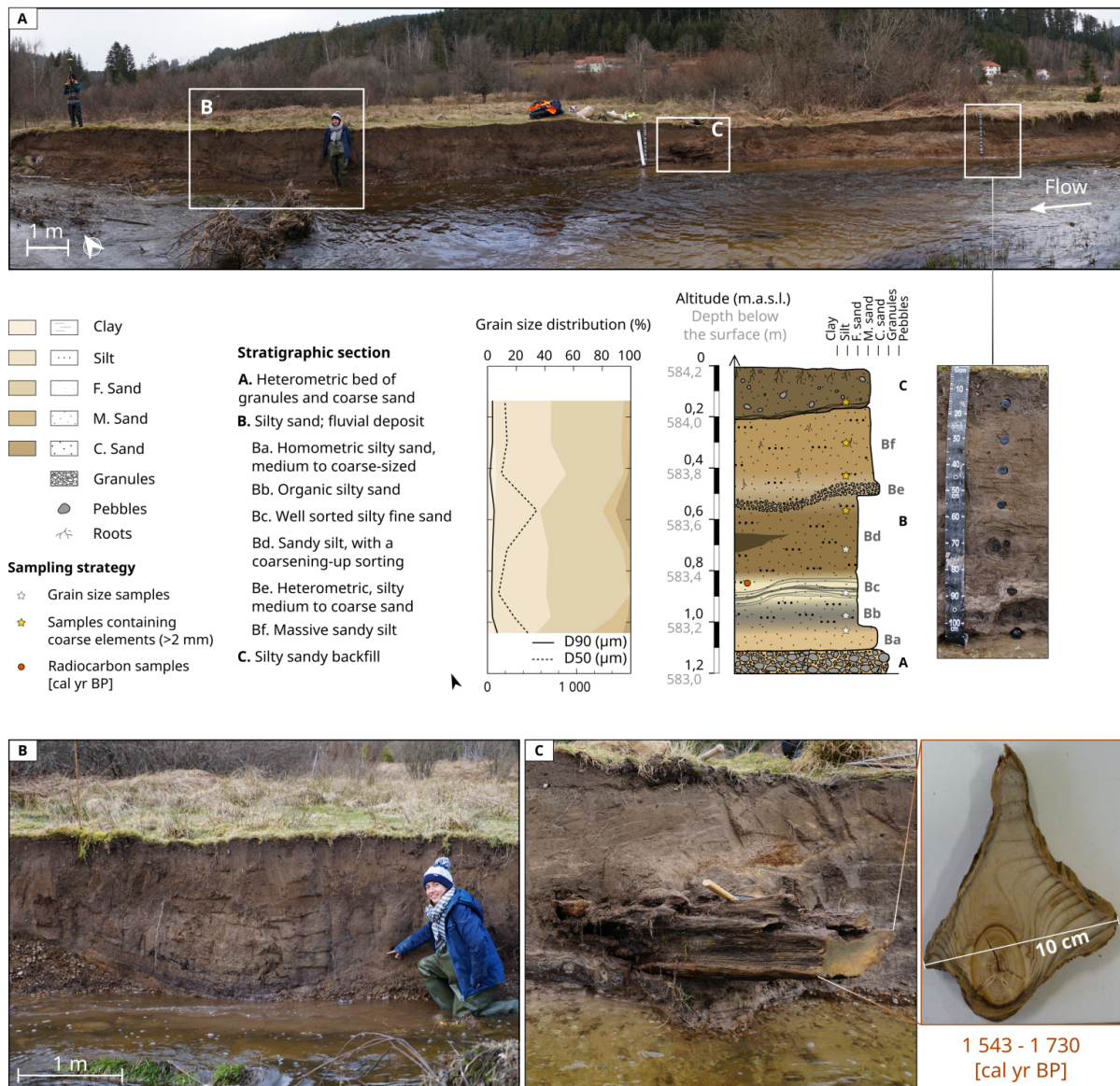


FIGURE 3.8: Log, grain size results and pictures of the outcrop RB located at Rain-Brice. (A) Right bank of the Cleurie. (B) Detail of the bank: a fine, dark and curved structure appears in a coarser environment. (C) Detail of the bank: a wooden trunk emerging from the deposits. Trunk sampled for radiocarbon dating.

The grain-size analysis reveals a profile with a low variation of the particle's diameter: mainly made of silt and fine sand, the D_{50} is about $74 \mu\text{m}$ and the D_{90} about $200 \mu\text{m}$ all along the

profile. Radiocarbon dating of the tree trunk provides an age-estimate for the enclosing sandy-unit, which is interpreted as composed of fluvial sediments, of ca. 1640 cal. BP (Fig. 3.8, Table 3.1).

3.5 Discussion: strong and multi-faceted glacial legacy in the Cleurie catchment implies post-glacial fluvial disequilibrium

The peculiarity of the Cleurie catchment was pointed out very early at the Vosgian scale (e.g. Hogard, 1840, 1842; Meyer, 1913; Nordon, 1931; Capot-Rey, 1938; Schouler, 1942). These studies, however, primarily focused on abundant glacial imprints in the landscape and so did the later ones (Seret, 1967; Salome, 1968; Flageollet et al., 1971). One must underline that a former glacial setting where three distinct glaciers spatially converged into a single valley, as it was most probably the case in the Cleurie, is unique within the Vosges. This was an uncommon phenomenon even at the scale of the Alps during the last glacial cycle (Pomper et al., 2017). Despite the wealth of geomorphological studies and some attempts of landscape reconstruction based on relative chronostratigraphy, (e.g. Flageollet, 1988), post-glacial evolution of the entire Cleurie catchment was almost completely left aside. Here, we first discuss how the multi-faceted glacial legacy has impacted the Cleurie landscape following deglaciation, specifically focusing on how it has affected various fluvial features and processes, eventually resulting in a state of pronounced geomorphological disequilibrium (Mason and Polvi, 2023).

3.5.1 The 'beheaded' Cleurie catchment: first key element explaining disequilibrium

The current spatial decoupling between the Vosgian main divide and the Cleurie catchment must be first discussed (Fig. 3.1A). Introduction lines of Flageollet et al.'s (1971) comprehensive study directly alludes to it: 'Modeste affluent de la Moselle, la Cleurie prend sa source à l'aval de Gérardmer, en plein cours d'une vallée beaucoup trop large pour elle [...]. Cette anomalie fut assez tôt remarquée et valut à la vallée de la Cleurie et à la haute vallée de la Vologne qui la précède en amont de Gérardmer une certaine renommée qu'attestent plusieurs études'. The end moraine located to the west of the Gerardmer Lake was probably formed by a E-W flowing glacier (Figs. 3.1B,C and 3.5), referred to as the former 'Vologne' glacier (Seret, 1967). The glacial lake, instead of spilling out through a breach in this 'pristine' moraine towards the west and the Cleurie valley, is drained by a small-dimensioned, NE-flowing tributary of the current Vologne river (i.e. the Jamagne). The confluence is located several kilometres NE of the Cleurie catchment, directly upstream of the Vologne straight

reach where it cuts deeply through Paleozoic granites along a main SE-NW orientation (i.e. Kichompré gorges; Fig. 3.1B).

The Cleurie catchment was formerly connected to the Vosgian main divide through the Vologne glacier fed by well-developed west-facing cirque(s) carved at the top of the main ridge. This glacier successively formed the Bas-Beillard and Gerardmer moraine complexes in the upstream part of the catchment, probably during the last glacial cycle (Fig. 3.5B; Flageollet, 1988). However, geomorphological decoupling linked to the Gerardmer's Lake drainage resulted in this post-glacial hydrographic setting which has two main implications. First, it explains why, despite the wealth of glacial landforms spread throughout the catchment, (i) erosional landforms, cirques in particular, are virtually absent and (ii) depositional landforms (i.e. moraines, tills, erratics...) largely prevail and represent, together with fluvio-glacial deposits, over 40% of the catchment area (Jautzy et al., 2024b). Let's remind here that the two other valley glaciers, i.e. Moselotte and Cellet, flowed upslope the Cleurie valley or bypassed the catchment's southern ridge from the Bouchot's valley, respectively (Fig. 3.1B). Second, this decoupling also led to the 'beheading' of a substantial part of the upper catchment, i.e. ~ 60 km². The Cleurie's headwaters nowadays that are the main stem's reach #1 and the tributary #3 (Figs. 3.4;3.5) merely are narrow step-pool systems (Fig. 3.9) characterised by very steep longitudinal slopes and small drainage areas (i.e. $>1\%$ and ~ 2 km² for reach #1; Fig. 3.5A), hence quite low water discharges. Flageollet et al.'s (1971) description above matches the tributary #2 currently draining the main former glacial valley (Fig. 3.4). This beheading process, at the expense of the Cleurie and the benefit of the Vologne, resulted in a drastic decrease of specific stream power (Yang et al., 2015; Yaaqoub and Essaifi, 2023), hence a depleted ability for the hydrographic network to fluvially reshape the landscape, at least in the upper part of the catchment.

3.5.2 Knickpoints (-zones) and variety of long profiles also reflect disequilibrium

3.5.2.1 Main trunk

At first glance, it is obvious the Cleurie's long profile bears several irregularities, hence strongly points to disequilibrium, but two key points must be discussed here. First, they are spatially distributed over the whole profile as they occur in almost all reaches of the river. Second, the four distinct shifts in linear arrays of slope-area data (Fig. 3.6A), i.e. two knickpoints and two knickzones, potentially indicate different geomorphological origins of these irregularities. Let's consider the knickpoints first. The upper one occurs in reach #1 and is a smooth slope break well matching the transition between granite to leucogranite on the southern Cleurie's valley wall in the headwaters (Fig. 3.4). Interestingly, a comparable smooth knickpoint is observed in the tributary #3 at a similar position along its profile (Fig. 3.6B). This



FIGURE 3.9: Picture of the Cleurie riverbed at the head of the catchment (#1; Fig. 3.4), corresponding to a narrow (~ 1.5 m-width) step-pool channel.

could highlight a lithological origin for these knickpoints as it was abundantly reported in other settings (e.g. Rixhon, 2010). The knickpoint occurring in the Cleurie's lowermost reach, by contrast, epitomises a sharp slope break at the valley outlet (Fig. 3.6B). It matches the sole bedrock occurrence across the whole valley floor where the active channel encounters a massive leucogranite outcrop and builds a ~ 6 m high waterfall, i.e. the so-called 'Saut de la Cuve' (Fig. 3.10A). This sharp slope break in such a downstream position strongly suggests important base-level variations in the confluence area (Beckers et al., 2015). Interestingly, a similar knickpoint is also observed in the downstream section of another Moselotte tributary, located

several kilometres upstream of the Cleurie confluence (i.e. the so-called Bouchot). These observations point to a general base-level drop linked to an incision episode of the Moselotte subsequent to the last glacial cycle that could explain a pulse of backward erosion propagating upstream into its tributaries. This is well in line with infra-red stimulated luminescence ages dating fluvial terrace deposits of the Moselotte located some kilometres downstream of the confluence around 19-22 ka, i.e. the last glacial maximum (Fig. 3.1B; Cordier, 2015). Importantly, whereas the Moselotte was able to cut through similar granite outcrops in the same area and form a two kilometres long gorge reach (Fig. 3.10B), the occurrence of this sharp knickpoint, i.e. waterfall, in the very downstream part of the Cleurie valley (and the Bouchot) means that the diffusion of the erosion signal further upstream must have been strongly buffered there. In this respect, Crosby et al.'s (2007) models propose a knickpoint forming (i) either at the catchment outlet and retreating upstream at a rate that is a power law function of drainage area or (ii) near a threshold drainage area, under which the channel's incision efficiency is hampered. Contrast in incision behaviour between the Moselotte (258 km²) and the Cleurie (78 km²) as well as beheading of the catchment's upper part that should have occurred approximately at the same time, i.e. after the last glacial maximum (assumed to be coeval with the formation of the Gerardmer Lake), could both argue for the second scenario.

As for the two ~0.5 and 2 km long knickzones, they occur in the second and third reaches, respectively. Whereas the former well matches the valley section where the predominant straight channel cuts through the Bas-Beillard moraine, the latter represents two thirds of the intermediate Cleurie reach (#3) where the meandering channel cuts through Le Tholy's morainic complex (Fig. 3.6B). Valley-floor narrowing leading to a confined river channel is clearly observed in both cases as well. As these knickzones are obviously associated with the occurrence of the most prominent depositional glacial landforms along the Cleurie valley (Figs. 3.5 and 3.6A), one can assume at this stage the primary control exerted by the geomorphological setting inherited (presumably) from the last glacial cycle on post-glacial fluvial adjustment. This will be discussed in more detail in section 3.5.3 .

3.5.2.2 Tributaries

Wide morphometric variability according to Wobus et al. (2006) is attested in the Cleurie catchment as all categories of river long profiles are represented: adjusted, straight, with one -or several- knickpoints and hanging valleys (Fig. 3.6B). A preliminary subdivision among hanging valleys can be operated based on the degree or intensity of slope break in the tributary's downstream section: it is either slightly (e.g. #8 or 9) or significantly (e.g. #5 or 11) perched above the main stem. If two main mechanisms are responsible for hanging valley formation, it primarily depends on what happens in the main stem or in the tributary (Crosby et al., 2007). The first one involves a quick incision pulse along the main trunk that overpasses

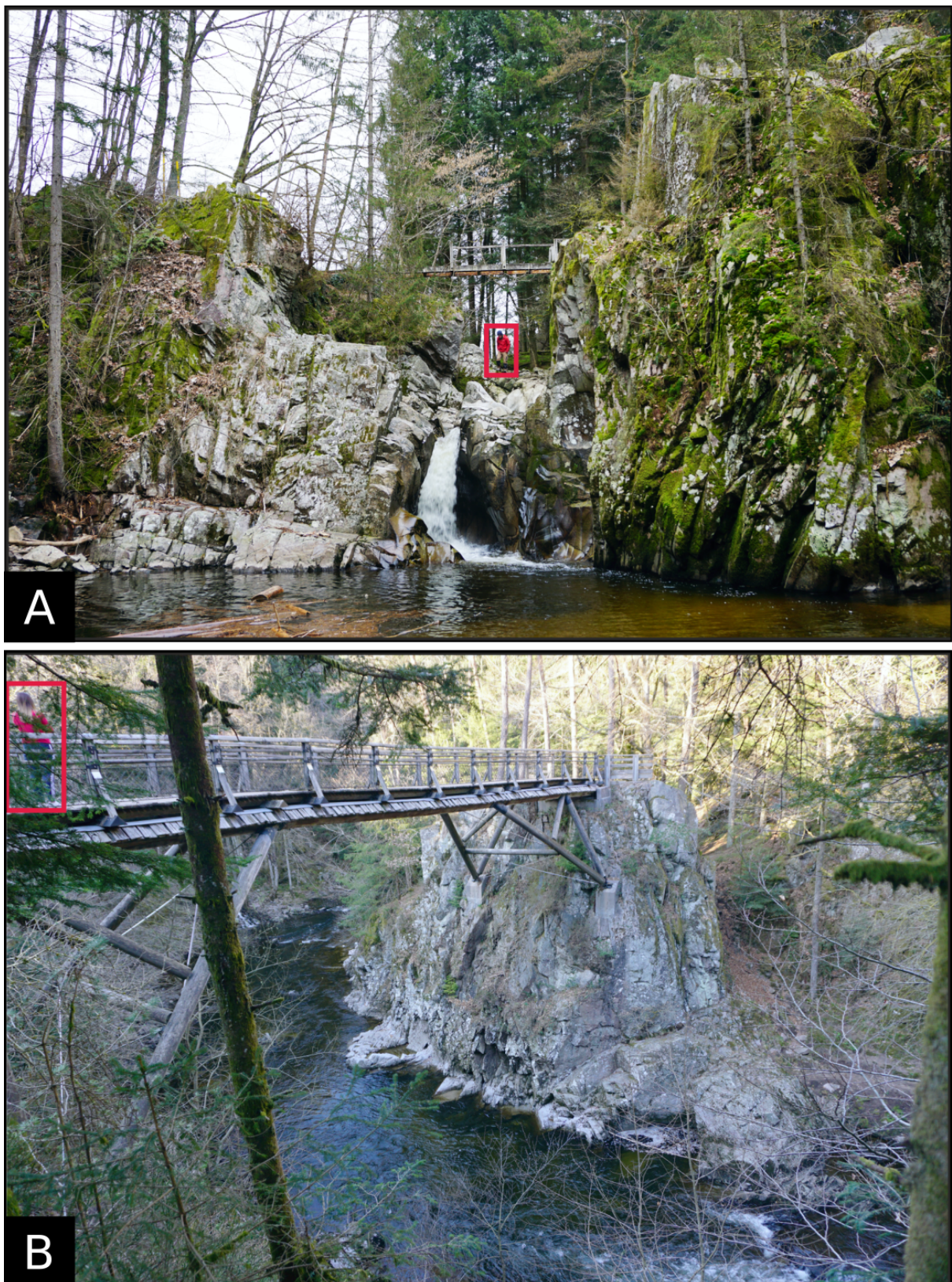


FIGURE 3.10: Pictures of the (A) ~ 6 m-high knickpoint ('Saut de la Cuve') occurring in the Cleurie's lowermost reach and the (B) ~ 2 km-long 'Gorges de Crosery' on the Moselotte river, occurring at ~ 6 km upstream of the confluence with the Cleurie river. The red rectangle is a human scale (~ 1.72 m).

a slope threshold at tributary junctions ('oversteepened tributary') whereas the second one relies on the tributary's hindered ability to incise due to a reduced sediment influx during the incision pulse along the main trunk ('outpaced tributary'). Whether the Cleurie's hanging tributaries were formed by the first or second mechanism -or both- requires, however, further investigations. In any case, starting from the Cleurie/Moselotte confluence, the degree of 'catchment disturbance' seems to reflect a spatial pattern as it apparently tends to increase upstream (Fig. 3.6B). On the one hand, the five downstream tributaries (#13-17) six to seven kilometres upstream of the confluence (roughly matching the Cleurie's #5 reach) display adjusted or straight profiles. On the other hand, tributaries' long profiles with one or several slope break(s) and hanging valleys predominantly occur in the catchment's upper half where the spatial clustering of glacial landforms is, by far, the highest. To sum up, one can assume a fluvial 'regulation' signal starting to retreat upstream along the main stem at a certain time after the last glacial maximum. However, it did not propagate throughout the entire river network, possibly partly buffered at the 6-m high 'Saut de la Cuve' waterfall (if coeval), and this could be related to the intensity of glacial overprinting in the landscape.

3.5.3 Peculiar spatial evolution of the bed-surface grain-size bears witness to strong glacial imprints in the landscape

In many fluvial systems worldwide, abrasion and hydraulic sorting lead to downstream gradation in bed material, i.e. downstream decrease in size related to decreasing longitudinal slopes (e.g. Houbrechts et al., 2011; Fryirs and Brierley, 2013). However, along-stream variations of the Cleurie's bed-surface grain-size tend to show an opposite trend (Fig. 3.6C). The lowest D_{50} (<50 mm) and D_{90} (≤ 80 mm) values are consistently observed upstream (except site #1 which is characteristic of the headwaters' step-pool system), especially along reach #2, whereas these parameters in the whole downstream section (from site VI to XIV) consistently remain much higher (≥ 75 mm and ≥ 190 mm, respectively). Two main points must be discussed here.

First, this unusual trend can be, again, largely interpreted as a twofold—both direct and indirect—glacial legacy. As for the direct impact, the three east-facing end moraines upstream of Le Tholy's morainic complex represent the main glacial feature in terms of topographic imprint given their (i) height above the Cleurie's valley floor (i.e. > 50 m; Fig. 3.5B) and (ii) volume of coarse material stored directly aside the active channel. In parallel, the sudden increase in D_{50} and D_{90} between sites V and VI perhaps is the most striking feature in terms of grain-size changes (Fig. 3.6C). Though only 0.2 km distant from each other, site V is located directly upstream of the uppermost Le Tholy's morainic complex whereas the next measurement matches its crossing (Fig. 3.4). Boulder occurrence reflects this as well: absence in site V versus consistent presence along the reach crossing the three successive moraines (i.e. 5

to 8 boulders in sites VI to VIII; Fig. 3.5C). If Hogard (1840) already identified this incised reach (Fig. 3.2A), steady increase in longitudinal slope forming the main knickzone along the active channel is also highlighted there (section, 5.2.1.; Fig. 3.5A-B). All of this strongly points to a local but massive input of coarse to very coarse particles due to channel's active lateral erosion that has substantially reworked moraine materials. An historical map interestingly reports significant gully erosion taking place in this moraine complex as a consequence of a massive flood event ('déluge de Sainte Anne' ; Salmon, 1770). Abundant gullies eroding the middle and the lowermost moraines as well as the intermediate area are possible geomorphological features of this event visible on LiDAR data (Fig. 3.4). Whether the fairly coarse and almost stable bed-surface grain-size documented downstream of site VIII to the valley outlet (Fig. 3.5C) results from (i) downstream mobilisation of large clasts or (ii) more local inputs originating from till material left behind by the Moselotte glacier in the Cleurie valley remains unresolved yet. Continuous downstream occurrence of boulders would rather argue for the second option as their size obviously exceeds the channel's ability to mobilise them. This is well in line with Bunte and Abt's (2001) 'relict or non-fluvial streams' resulting from channel incision into underlying glacial deposits (e.g. till), thereby unearthing large boulders that frequently are of untransportable size.

As for the indirect impact, the lowest D_{50} and D_{90} (sites II and IV) remarkably match the former extension of the Beillard palaeolake (Fig. 3.4). It also well matches the Cleurie reach #2 where the meandering channel flows through a up to 1 km wide valley floor (Fig. 3.6B). Absence of coarse material reworked from major glacial landforms combined to full decoupling between channel and hillslopes (preventing any lateral inputs of coarse material subsequent to deglaciation) both explain this unusual pattern.

Second, measurements in tributaries #5 and 11 directly upstream of their confluence were carried out to check their input material. On the one hand, D_{50} and D_{90} of tributary #11 (site IX) are pretty much in the same range as those of the main stem measured between sites VI to XIV (Fig. 3.5C). Thus, one can not assess whether this tributary has an impact in terms of bedload coarsening or fining in the Cleurie reach #4. On the other hand, D_{50} and D_{90} of tributary #5 are substantially higher than those measured in sites II and IV along the main stem (reach #2), located up- and downstream of the confluence, respectively (Fig. 3.5C). Interestingly, the lack of any significant bedload coarsening downstream in the Cleurie's active channel points to its inability to mobilise bedload material with a D_{50} amounting to 90 mm in this reach. One must note here the decrease in longitudinal slope linked to the presence of the former palaeolake(s). In any case, beyond the spatial link between glacial landforms and along-stream variations of the bed-surface grain-size and its unusual pattern, this observation further argues for a fluvial disequilibrium along the main stem.

3.5.4 Reconstruction of the formation stages of glacial-related and fluvial units, in relation with the successive palaeolake levels.

The updated geomorphological mapping (Fig. 3.5) includes (1) previously mapped glacial-related units located at high elevations along the hillslopes of the main stem and some tributaries, and (2) previously unmapped fluvial morpho-sedimentary units covering the valley floor. The glacial-related units identified by Flageollet et al. (1971) are interpreted as glacio-lacustrine deposits resulting from the gradual lowering of the former paleolake (lower, intermediate, upper, and Bas-Beillard levels) and fluvio-glacial terraces ('Lower Cleurie terrace'). The fluvial units are interpreted as the modern floodplain and a low fluvial terrace. Figure 3.11 plots the absolute elevation of each identified units (both glacial-related and fluvial), in relation with the long profiles of the Cleurie main stem and the morainic complexes. Put together, Figures 3.5 and 3.11 help locate and understand the different morpho-sedimentary units in the landscape, illustrating key events such as successive lake levels, glaciers retreat, and the formation of fluvio-glacial and alluvial terraces. This sequence of events demonstrates the complex interplay between glacial and fluvial processes that have shaped the valley over time.

According to Flageollet et al. (1971) and Flageollet (1988), the successive lake levels, as depicted in Figures 3.5 and 3.11, include the upper, intermediate, Bas-Beillard, and lower levels. Based on the authors' observations and our geomorphological mapping (Fig. 3.5), we can propose a summary of the formation stages of the glacial-related and fluvial units.

- **Upper level:** This level corresponds to that of a lake that was contained to the east by the Vologne glacier, in the vicinity of the present-day Gérardmer moraine, and to the west by the Moselotte glacier, whose altitude was higher than that of the Tholy's morainic complex. The deltaic deposits present in the Cellet valley indicate that the banks of the lake were also established there. The outlet was the Rechaucourt pass.
- **Intermediate and Bas-Beillard levels:** These two levels correspond to two successive levels of stability of the lake, which has seen its level gradually lowered as the ice melted. The deposits associated with the intermediate and Bas-Beillard levels extend from the Costet to the Tholy's morainic complex, then from the Bas-Beillard to the Tholy's morainic complex, respectively. At these altitudes, the Rechaucourt pass could no longer serve as an outlet, and it is likely that the meltwater flowed between the hillslopes and the Moselotte glacier, which must still have extended as far as Le Tholy.
- **Lower level and fluvio-glacial terraces in the lower Cleurie valley:** With the continued retreat of the Moselotte glacier, a fourth and final level of lake stability can be assumed, forming the lower level deposits. These are located exclusively in reach 4, downstream

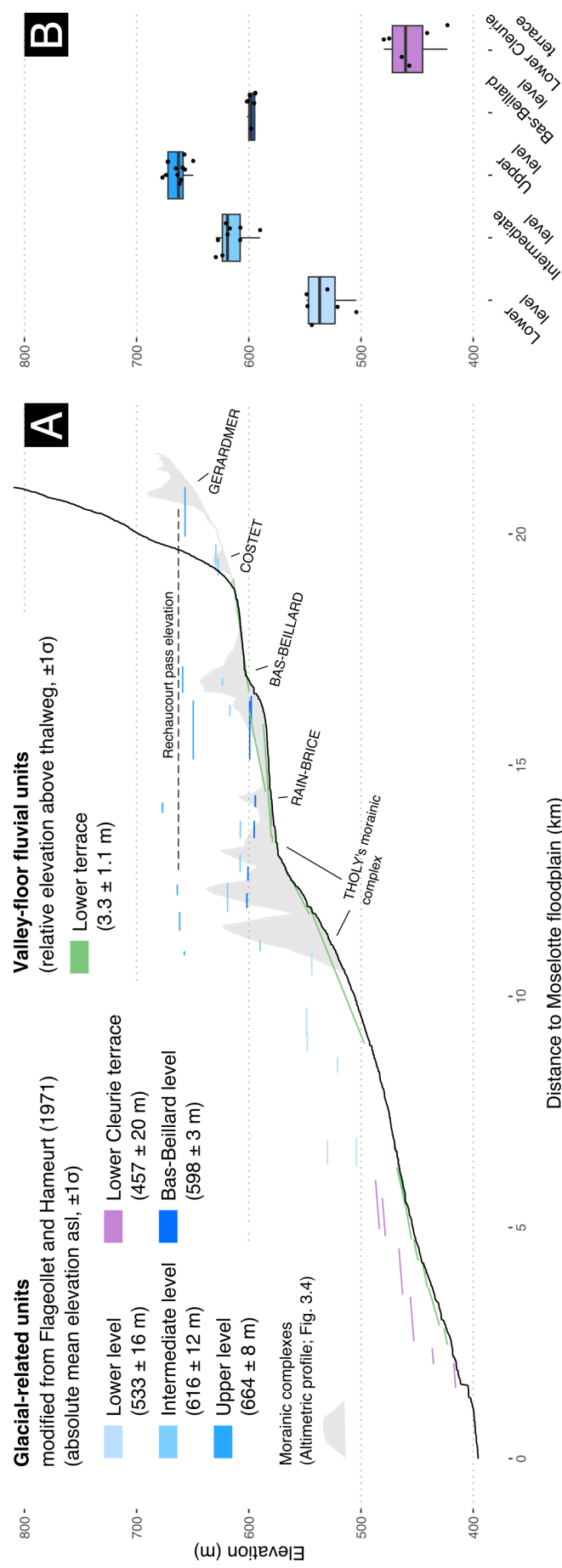


FIGURE 3.11: (A) Longitudinal profile of the main channel of the Cleurie, showing the absolute elevation of units mapped by Flageollet et al. (1971) and those of the valley-floor units mapped in this study. The elevations of the former correspond to the average elevations of each unit extracted from the 1 m DEM. For the lower terrace, elevation values were extracted upstream and downstream of each unit to represent their slope. (B) Distribution of average elevation values by type of unit mapped by Flageollet et al. (1971).

of the Tholy's morainic complex. We can also assume that the terraces of the 'Lower Cleurie' (reach 5; Fig. 3.5A) were formed at the same time, by the juxtaglacial flow of meltwater from the tongue of the Moselotte glacier.

Finally, after the complete retreat of the Moselotte glacier tongue, the Cleurie river was able to incise the Tholy's morainic complex. The lower part of the valley then began to readjust, leading to the formation of alluvial terraces. The lower terrace reflects a phase of valley-floor stabilisation and subsequent erosion, leading to the formation of the modern floodplain.

3.5.5 Palaeoenvironmental information derived from the sedimentary sequences and tentative chronological framework

The profile of Morte Femme is composed of three main sedimentary units overlying each other (Fig. 3.7). The lower unit consists in flat, clayey-silty laminae that are interpreted as being deposited at the bottom of a lake, with an alternation in environmental conditions leading to slight changes in the sediment type. The paleolake would have been fed mostly with minerogenic input, since organic matter content is, from visual observations, very reduced in the sediments. Some organic material was however found in the sediment upon sieving, which allowed for radiocarbon dating: fibrous matter that was tentatively interpreted as coming from fragments of sedges or grasses, and charcoal particles. Isolated pebbles in the decimetric range were found embedded into the laminae (Fig. 3.7G). These could be interpreted as dropstones, although a contemporaneous deformation of underlying laminae should be present, which is not clear from field observations (due to the plastic behaviour of the clayey sediments during sequence cleaning). Noticed elements point towards a periglacial environment in the direct vicinity of the MF outcrop, although reduced vegetation may have been present in the surroundings, e.g. sedges/grasses growing at the edge of the paleolake. Charcoal fragments $>250\ \mu\text{m}$ potentially suggest higher vegetation forms (e.g. shrubs or trees) in the catchment, and ambient temperatures sufficiently high at times to sustain fires. Contamination of the studied laminae with organic matter, e.g. through rootlets intrusion or during the sampling process, cannot be completely excluded, although (i) great care was taken during subsampling to remove exposed edges and (ii) the clayey-silty nature of the sediments renders it difficult for plants to penetrate it. It should be noted, additionally, that no sign of bioturbation was observed in the laminae either in the field or during lab processing. The three dates obtained from the lacustrine rhythmities are in stratigraphic order and range from 10575-10281 to 9528-9294 cal. BP (1281 - 753 years) over 90 cm (Table 3.1). This would amount to a mean deposition rate of 0.7 to 1.2 mm per year, which doesn't correspond fully to a preliminary visual examination of the laminae, which seem to be pluri-millimetric in size (Fig. 3.7), and therefore tend to exclude - if the dating is right - that a continuous, seasonal

signal is registered in these alternations. Observations of thin sections of the laminae under the microscope, as well as pollen investigations of individual laminae, may help resolve this question. The middle unit comes at a discordance to the lower one, with a clear erosion pattern at the boundary (Fig. 3.7). This consists of pebbly and sandy units, interpreted as fluvial deposits cutting through the underlying lacustrine clayey-silty sediments. It contains centimetric wood fragments (including *Pinus*; Table 3.1). Three dates obtained on such fragments, at ca. 13'190, 13'130 and 11'610 cal. BP, attest the presence of tree vegetation in the Cleurie catchment, at or upstream the section, at these times. The dates are, however, inconsistent with those determined in the underlying lacustrine unit. Their stratigraphic inversion within the unit points to remobilisation of older sediments by the river upstream, prior to their redeposition at the sequence. These two dates should, therefore, be considered as giving a maximum age for the fluvial deposits at Morte Femme, therefore corresponding to the establishment of the Lower terrace in the valley (Fig. 3.5). The organic-rich upper unit caps the middle unit and suggests the colonisation of the sequence's location by peatland following lateral migration of the river. Radiocarbon dates are inconsistent in this unit, which may be related to easy (woody) root penetration into organic sediments. Individual bark or wood fragments selected in the unit for dating (Table 3.1) could therefore correspond to root parts and should be considered as providing a minimum age. The older of the ages obtained in the peat, at ca. 8'600 cal. BP is, however, compatible with all ages found in underlying units. The younger age, at ca. 550 cal. BP, suggest persistence of the peaty unit sedimentation to more or less modern times. The tree trunk present in the RB sequence (Fig. 3.8) suggest the establishment of the floodplain (Fig. 3.5B) around ca. 1640 cal. BP, since remobilisation of so large a piece of wood (Fig. 3.8C) seems extremely unlikely.

Overall, two scenarios can be derived from the radiocarbon dating results from the two sequences of Morte Femme and Rain Brice. A lake, with mostly minerogenic input, seems to have persisted above an elevation of 606 m (Fig. 3.7) at least up to ca. 13'190 cal. BP (oldest date in the overlying fluvial deposits). If the three dates in chronological order in the lacustrine laminae are considered correct, they even sign the persistence of this lake well into the Holocene. This lake could be a remnant of the 'intermediate level' stage described by Flageolet et al. (1971), and owing its existence to the extensive moraine ridge present downstream of the Morte-Femme area. A mostly mineralogical content is frequently observed in Late Glacial lakes in the Vosges area, but are highly unusual in the region for 13 ka BP or even more 9 ka BP, when most of wetland sediments in the Vosges are richer in organic matter (de Klerk, 2014). Such sedimentation is found nowadays in Alpine regions. This may reflect very particular conditions at Morte Femme during the Late Pleistocene-Early Holocene with even potentially the remnant of ice in the surroundings, inducing a micro-climate. Part of the catchment at, or upstream of, Morte-Femme should have been ice free with some tree development at ca. 13'190-13'130 and 11'610 cal. BP. It should be noted that the earlier date found

in the peat sediments at ca. 8'600 cal. BP is corroborated by the literature: a date of 6 600 $\pm$ 370 y BC is mentioned in the Bruyères sheet of the BRGM geological map for the Bas Beillard peatland at 2 m depth (deeper organic layers). In general, the lake disappearance and fluvial incision (max 13'190 cal. BP, potentially between 9'410 and 8'600 cal. BP) at Morte Femme, and terrace build-up at Rain-Brice (around ca. 1640 cal. BP) are far more recent than expected, and could explain the strong fluvial disequilibrium of the Cleurie hydrographic network that we characterised in this study.

Chapitre 4

Quantifying the impact of human pressure exerted on three Vosgian rivers (Fecht, Moselotte, Vologne) over the last two centuries

Some of the data used in this chapter were produced as part of two Master research projects (Berthier–Laumond, 2022; Claudepierre, 2022). We plan to submit this article in the journal **Anthropocene**. Results were presented as a poster at **INQUA2023** (Rome, Italy) and **Quaternaire14** (Rennes, France), and as an oral at **EGU2024** (Vienna, Austria).

4.1 Introduction

Over the last centuries, the morphodynamic of fluvial systems in most regions of the world has been profoundly affected by anthropic pressure (Wohl, 2020; Surian, 2022; Brenna et al., 2024), which is now recognised as a major driver of modern fluvial changes (Liébault and Piégay, 2002; Salit et al., 2015; Cooper et al., 2018; Verstraeten et al., 2017). Anthropic pressure on fluvial systems refers to direct and/or indirect human interventions at the catchment, valley-bottom and channel scale, such as land use changes (Liébault and Piégay, 2002; Delile et al., 2016; Jautzy et al., 2021), flow regulation (Winterbottom, 2000), gravel mining (Comiti et al., 2011; Ziliani and Surian, 2012), as well as channelization, embankment, levee, and dam/weir construction (Arnaud et al., 2019; Mandarinò et al., 2019; Belletti et al., 2020). With only a third of rivers meeting the criteria for ‘good ecological status’ according to the Water Framework Directive, Europe probably has the most degraded rivers in the world (Belletti et al., 2020),

which is related to Europe's long history of river occupation and exploitation. European fluvial systems experienced increasing anthropic pressure at various scales, impacting both catchment-wide sedimentary processes and channel/valley morphodynamics across different historical periods (Rinaldi et al., 2013; Brown et al., 2018). Human pressure began as early as the Neolithic period with initial phases of deforestation linked to pastoral and agricultural activities (Williams, 2000). This trend intensified over time, with notable peaks during periods like the Roman era, culminating in extensive deforestation in the 19th century. Between the 17th and 19th centuries, increased runoff and erosion resulted in widespread channel aggradation, transforming many mountain and piedmont rivers into braided channels (Wyżga, 1993; Gurnell et al., 2009). With the onset of industrialisation and urbanisation, anthropic pressure on river systems intensified dramatically. At the catchment scale, afforestation efforts led to a decreasing sediment supply, followed by channel narrowing and incision (Lach and Wyżga, 2002; Keesstra et al., 2005). Multi-threaded channel patterns declined in many rivers (Gurnell et al., 2009). At the same time, human interventions also occurred at the channel and valley-bottom scales. They included damming, channelisation, and flow diversion primarily for flood control, agricultural irrigation, and navigation enhancement on major rivers (Billi and Rinaldi, 1997). Weirs were constructed along the headwaters of mountainous rivers and larger dams downstream, significantly reducing sediment supply to downstream areas and disrupting water and sediment fluxes. Finally, post-World War II urban and industrial growth had pronounced impacts on river morphology, including increased infrastructure development, pollution, and drastic alterations to riparian landscapes. Human interventions during this period focused on constructing larger dams, straightening channels to facilitate transportation and industry, and implementing flood control measures and water management strategies to meet growing industrial and urban demands.

Numerous studies have highlighted the link between the authorisation of European valley bottoms and channels and the fluvial changes measured during the late modern period (Ziliani and Surian, 2012; Salit et al., 2015; Arnaud et al., 2019). Because fluvial dynamics, including forms and associated processes, conditions habitats, biodiversity, and ecosystem services, any alterations of morpho-sedimentary processes or channel forms inevitably leads to habitat loss, reduced biodiversity and diminished ecosystem services (Brown et al., 2018). In recent decades, increasing recognition of human impact on the degradation of European river systems, and the awareness of the huge negative effects of this evolution, consequently led to a growing policy to restore fluvial forms and dynamics (García et al., 2021).

In this respect, one of the key target of the European Water Framework Directive adopted in 2000, is to restore water bodies, including rivers, to a 'good ecological state' (Schmitt et al., 2007; Dufour and Piégay, 2009; Maaß et al., 2021). Defined as a low deviation from 'reference conditions' (Gob et al., 2014), a 'good ecological state' might thus suggest a return to a 'natural' state (Comiti, 2012). However, the relevance of a return to a 'natural' state is widely

debated, as it is challenging to identify and not necessarily achievable, or even desirable, depending on current human occupation and needs (Brown et al., 2018). When considering the management/restoration of a fluvial system, rather than aiming for 'reference conditions', it has been shown that it is preferable to target the preservation or even restoration of fluvial ecosystem services (Dufour and Piégay, 2009; Wohl, 2020). This requires on the one hand to characterise the evolutionary trajectory of the fluvial system and on the other hand to assess the respective share of its various controlling factors over time, which conditions the possibilities of restoration (David et al., 2016).

However, because of the non-linear interactions between anthropic and natural factors, interpreting the potential cause-and-effect relationships in fluvial systems affected by multiple human interventions is a very challenging task (Downs et al., 2013; Brenna et al., 2024). Moreover, in a recent literature review conducted on studies devoted to human controls on fluvial changes, Downs and Piégay (2019) highlighted that most of the demonstrations concerning cause-and-effect relationships can often be considered to be not sufficiently robust or lacking convincing scientific arguments and precise quantification. First, human forcings, whether direct or indirect, are rarely precisely spatialised. Except in the case of very localised studies, which analyse the morpho-sedimentary impacts of a particular development, on a particular river reach. Second, the cumulative impact resulting from potential multiple anthropic drivers are typically overlooked. Third, the chronology of events is often only partial or rather vague (Verstraeten et al., 2017).

In this study, we focus on three adjacent catchments of the southern part of the Vosges Massif (NE France), the Fecht, Moselotte and Vologne catchments. The VM is the most densely populated massif in France (Fig. 4.1A) and one of the most densely populated in Europe (Rochel, 2004). It was significantly developed during industrial era (19th century) to produce the hydropower needed to run various industries (essentially textile and paper industries, sawmills) and to irrigate cultivated lands (Parmentier, 2019). As a result, of all French massifs, the vosgian rivers holds the record of the highest average number of obstacles to flow per km of streams (over one obstacle per km; Fig. 4.1B). The main goal of this study is to assess the extent to which the anthropisation of river channels, valley bottoms and catchments of the Fecht, Moselotte and Vologne has affected their fluvial morphodynamics over the last two centuries. We focus on this period for two reasons. First, it is the one over which we can gather the most information concerning human pressure. Second, in principle, it corresponds to the period when the greatest pressure occurred. As a secondary goal, we test the value of using documentary archives to assign construction periods to weirs still present in the river channels. We take advantage of documentary archives produced by the 'Ponts et Chaussées' administration. Stored in French departmental archives, they contain many requests for construction of hydraulic structures along French rivers, from the beginning of the 19th to the

end of the 20th (Payraastre, 2005). Although rarely used, few studies have successfully demonstrated the value of these specific archives in fluvial geohistorical reconstruction, focusing on the last two centuries (Jacob-Rousseau, 2015; Mesmin et al., 2024).

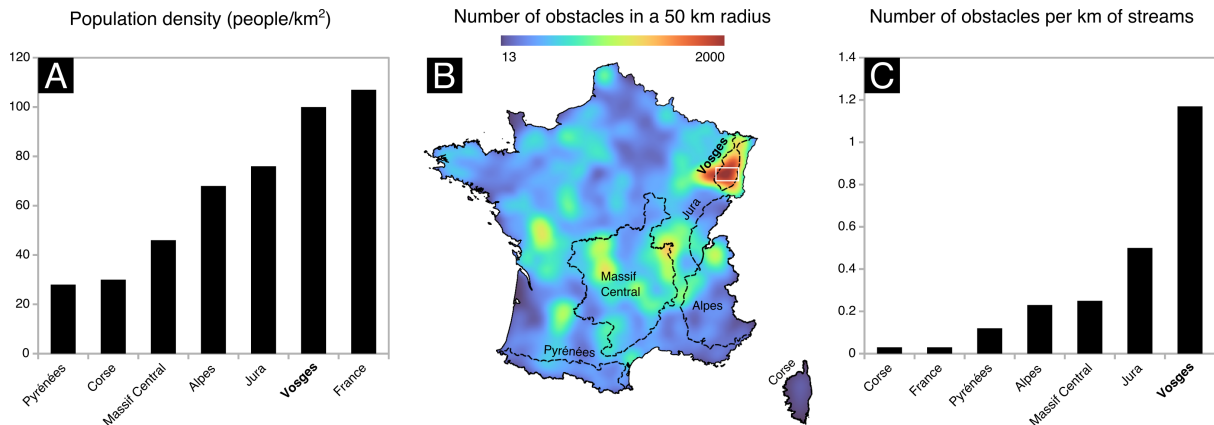


FIGURE 4.1: Comparison of (A) population density, (B) density of obstacles to flow and (C) number of obstacles to flow per kilometres of stream, in all French mountain ranges. Population data: Filosofi2017 (INSEE). Hydrographic network: BD Carthage (IGN). Obstacles to flow: ROE (2019). Mountain ranges are delineated according to the 'Observatoire des Territoires' (<https://www.observatoire-des-territoires.gouv.fr/kiosque/zonage-les-perimetres-de-massifs>). The white rectangle (B) locates the three studied catchments.

Over the last two centuries, (i) anthropogenic pressure, (ii) potential morpho-structural controls, and (iii) channel planform evolution were quantified along 500 m elementary reaches (ER) for each studied river, spanning their entire length. (i) At the catchment scale, the evolution and distribution of land use was reconstructed and analysed between the mid-19th century and the beginning of the 21st century, based on a digitised old map and a photo-interpreted orthophoto. At the valley-bottom and active channel(s) scales, the spatio-temporal evolution of the construction of weirs was reconstructed, by producing a database containing all the weirs currently and formerly present on the main channel of the three rivers. Documentary archives ('Ponts et Chaussées') were analysed, in order to assign *termini ante* or *post-quem* for the construction of the weirs. The total amount of paper documents analysed corresponds to approximately 17 m when stacked. A construction period was assigned to all the bridges currently present across the three rivers, using an old map (mid-19th) and two aerial photographs from the 20th and 21st centuries. All dykes currently present in the valley bottoms of the three rivers were mapped. (ii) The influence of morpho-structural controls was assessed by measuring the width of the valley floor, the presence of (fluvio-)glacial deposits in the valley-bottom, the slope of the talweg and the longitudinal distribution of the grain size of the bed surface. (iii) Finally, the diachronic planimetric evolution of the three rivers was reconstructed by digitising the active beds from the old map (mid-19th) and two aerial photographs from the 20th and 21st centuries.

First, the uncertainties and biases associated with the historical data used in this study are

discussed (documentary archives and maps). We also discuss the value of using documentary archives in this kind of approaches. Second, a geomorphological diagnosis of the three studied rivers is presented, and the evolutionary trajectory of channel patterns and fluvial dynamics over the last two centuries is discussed in relation to potential human and morpho-structural controlling factors. Finally, we argue that the results of this study might be usefully converted into management and restoration strategies, the most efficient and sustainable possible.

4.2 Study area

This study focuses on three rivers located in the southern part of the VM, which is the most densely populated French Massif (Fig. 4.1). The Fecht River is located on the eastern side of the massif and flows into the Ill, in the Upper Rhine Graben. The Moselotte and Vologne rivers are located on the western side of the massif and flow into the Moselle river, towards the Lorraine plateau (Figs. 4.2A; 4.3).

The Fecht River is 49 km long, has a catchment area of 520 km², a mean active channel width of 23 m and a mean channel slope of 8.2 ‰ (Table 4.1). The Fecht valley-bottom is particularly wide, especially in the downstream half (width ~1000 m). The Fecht ends its course on a large alluvial fan in the Upper Rhine Graben, at the outlet of the VM. The Fecht's bi-annual discharge is 61 m³/s. Mean annual precipitation is 1019 mm, with high spatial variability (± 302 mm). The lithology of the Fecht basin is mainly plutonic, with metamorphic outcrops upstream and sandstone remnants on the northern interfluvium. During the LGM, several glacial tongues descended into the valley, probably not exceeding the first third of its course (Andréoli et al., 2006). According to the BRGM geological map, glacial-related deposits cover 4% of the catchment area.

The Moselotte river is 47 km long, has a catchment area of 361 km², a mean active channel width of 19 m and a mean channel slope of 11.7 ‰ (Table 4.1). The Moselotte valley-bottom is steeply incised and relatively narrow over the upstream two-thirds (width ~100 m), then widens in the last third (width ~1000 m). For the last two kilometres, the Moselotte ends its course in the Moselle floodplain. The Moselotte's bi-annual discharge is 112 m³/s. Mean annual precipitation is 1356 mm, with low spatial variability (± 14 mm). The lithology of the Moselotte basin is largely plutonic. It is assumed that the Rissian glaciation covered the entire Moselotte watershed, while the Würmian glaciation was made up of several glacial tongues, including a remarkable diffuence at the confluence with the Cleurie (Seret, 1967). According to the BRGM geological map, glacial-related deposits cover 22% of the catchment area.

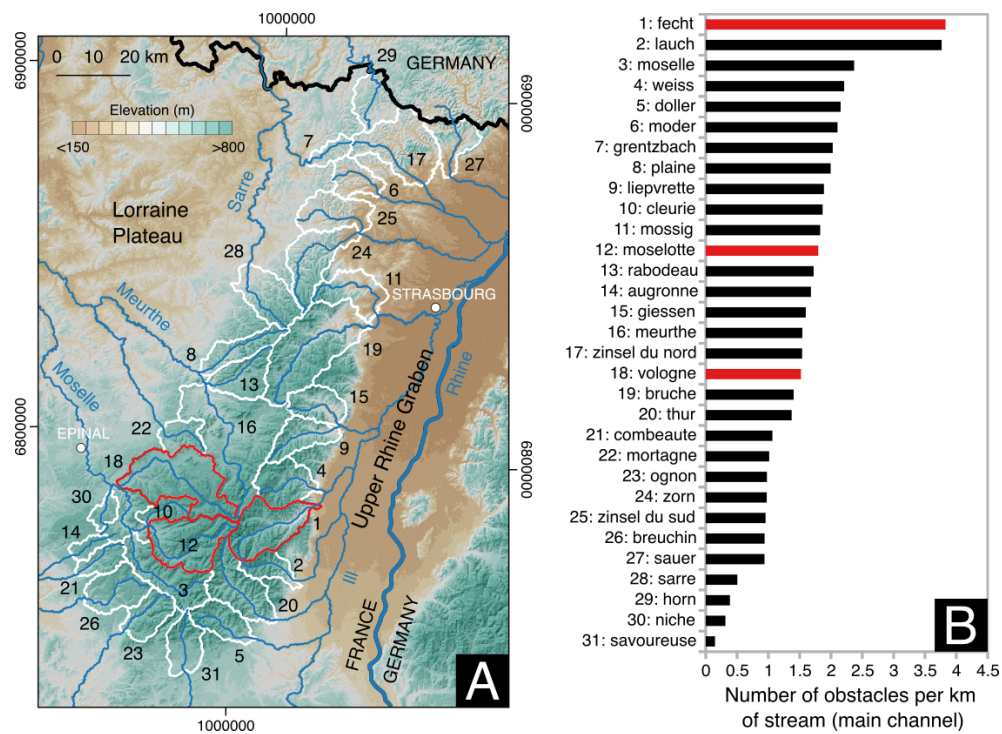


FIGURE 4.2: Comparison of the number of obstacles to flow per kilometres of river for every Vosgian catchments. Red catchments corresponds to the Fecht, Moselotte and Vologne. Hydrographic network: BD Carthage (IGN). Obstacles to flow: ROE (2019).

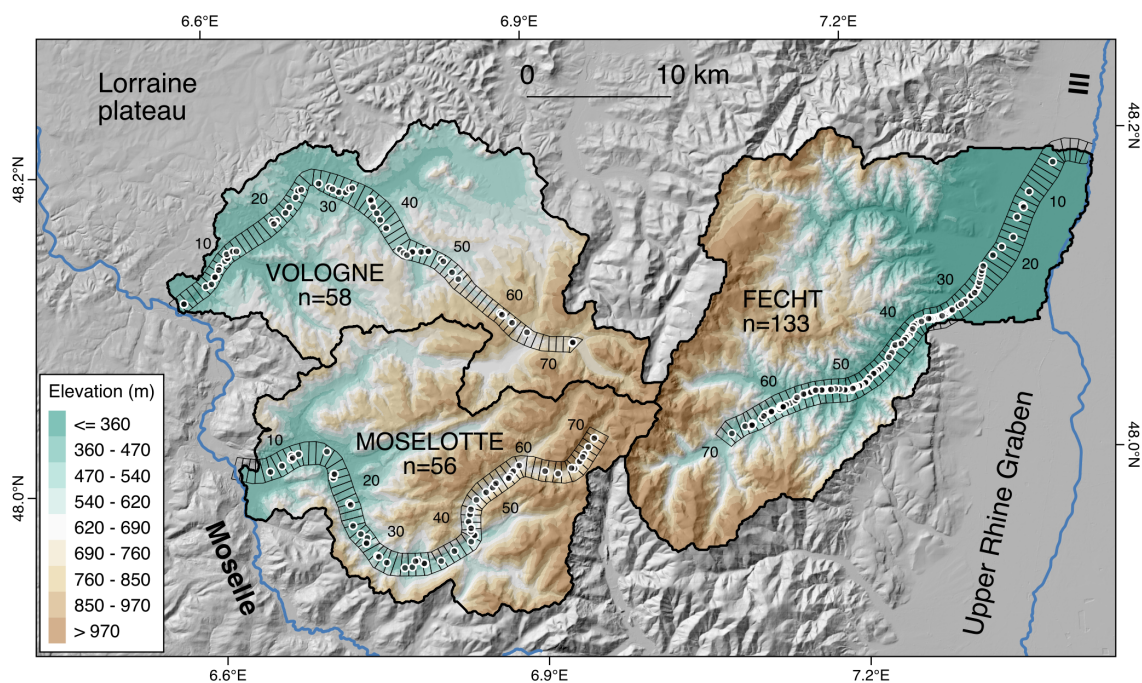


FIGURE 4.3: Topographical map of the Fecht, Moselotte and Vologne catchments. Location and number of obstacles to flow affecting the main channels of the Fecht, Moselotte and Vologne rivers, according to the French national database of obstacles to flow (ROE, 2019).

TABLE 4.1: Characteristics of the studied rivers and associated catchments. ^aThe length of the studied channel (see Fig. 4.4) was restricted by the visibility of the channel on the état-major map (see methodology). The total length of the channel goes up to the source of the river. ^bMean active channel width was computed on the linear studied. ^cMean active channel slope was computed on the linear studied. ^dQ₂ values were computed over the 1972-2024, 1993-2024 and 1988-2024 periods for the Fecht, Moselotte and Vologne, respectively (<https://www.hydro.eaufrance.fr/>), and scaled to the catchments areas. ^eMean annual precipitations were calculated in the catchments using the national hydroclimatic database of Météo-France (Vidal et al., 2010), over the 1958–2009 period. ^fPercentage of glacial-related deposits were extracted from BRGM’s geological map.

	Fecht	Moselotte	Vologne
^a (Total) Studied channel length (km)	(49) 38	(47) 43	(50) 43
Catchment area (km ²)	518	361	367
^b Mean active channel width (m)	23	19	18
^c Mean active channel slope (‰)	8.2	11.7	9.8
Q ₂ (m ³ /s) ^d	61	112	56
^e Mean annual precipitations (± sd) (mm)	1019 (± 302)	1356 (± 14)	1321 (± 99)
Mean elevation (± sd) (m)	611 (± 301)	751 (± 199)	636 (± 164)
Mean catchment slope (± sd) (°)	15 (± 10)	14 (± 8)	10 (± 7)
Main lithologies	Plutonic / Metamorphic	Plutonic	Plutonic / Metamorphic / Sandstone
^f Percentage of glacial-related deposits (%)	4	22	14

The Vologne river is 50 km long, has a catchment area of 367 km², a mean active channel width of 18 m and a mean channel slope of 9.8 ‰ (Table 4.1). The Vologne valley-bottom is steeply incised and relatively narrow in the upstream third (width ~100 m), then widens considerably in the second third (width ~1000 m) and finally narrows again (width ~500 m) before its confluence with the Moselle. The Vologne’s bi-annual discharge is 56 m³/s. Mean annual precipitation is 1321 mm, with moderate spatial variability (± 99 mm). The lithology of the Vologne basin is relatively varied. It is essentially plutonic upstream, metamorphic in the middle and mainly sandstone downstream. The Vologne basin is thought to have been only partially glaciated (Seret, 1967; Flageollet, 2002). Former glaciations would not have extended much further than the upstream half of the catchment. According to the BRGM geological map, glacial-related deposits cover 14% of the catchment area.

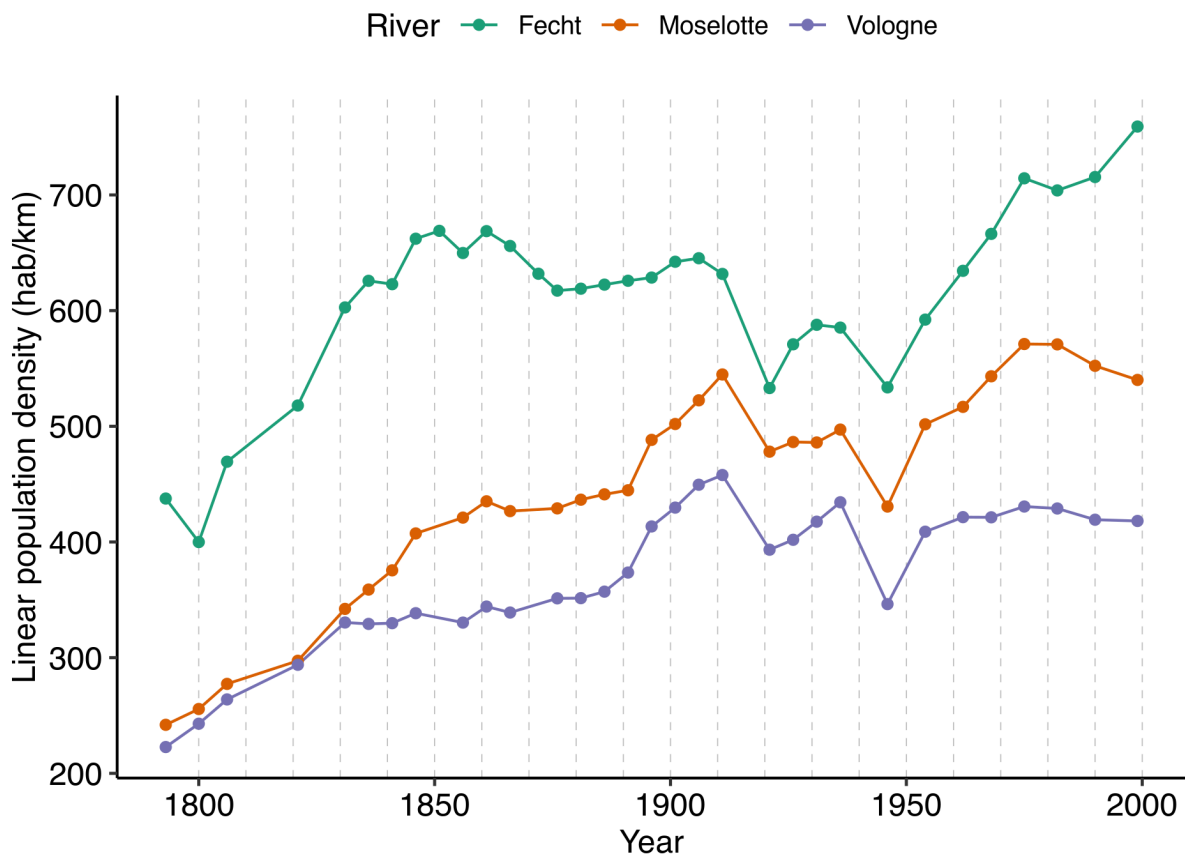


FIGURE 4.4: Demographic evolution of the Fecht, Moselotte and Vologne valley-bottoms, from 1793 to 1999. The linear population density corresponds to the population of all the municipalities crossed by the rivers, divided by their total length. Source: BD CASSINI (Ruas et al., 2014).

The demographic evolution of the Fecht, Moselotte and Vologne valleys over the last two centuries is directly linked to their industrialisation and to various historical events. It explains the human pressure on the channels and valley-bottoms. Figure 4.4 gives an overview of the human pressure trajectory of each valley, through the population density of their valley-bottoms per main channel length. From the beginning of the 19th century to WW1, the population density of the Fecht, Moselotte and Vologne valley-bottoms increased by around 60%, 110% and 90% respectively. Trajectories differ, however, between the Moselotte and Vologne (west side) and the Fecht (east side). The population density in the valley bottoms of the Moselotte and Vologne showed a consistent increase until the mid-19th century, followed by a period of relative stability until the end of the 19th century, and then a sharp rise into the early 20th century. In contrast, the Fecht experienced a continuous and more pronounced increase until the mid-19th century, followed by a significant decline between 1860 and 1880. This decrease can be related to the German annexion of the eastern side of the massif in 1871, following the Franco-Prussian war (Roth, 2014). Finally, from the second half of the 20th century onwards (post WW2), all three valleys experienced demographic growth, the most significant of which corresponds to the Fecht valley, followed by the Moselotte and finally

the Vologne, whose demographics remained almost constant to nowadays. These differences in trajectory and intensity of human pressure of the three valleys, might partly explain the anthropisation gradient measurable through the quantity of flow obstacles currently present in the three rivers.

According to the French national database of obstacles to flow, the Fecht river has the highest linear density of flow obstacles in the entire Vosges Massif ($\sim 4/\text{km}$, 133 in total; Figs. 4.2B; 4.3). The Moselotte and Vologne rivers have an obstacle density that is relatively average for the Vosges Massif, with 1.8 and 1.5 obstacles per kilometre, respectively. They currently have a total of 56 and 58 obstacles in their main channel (Fig. 4.3).

4.3 Data and methodology

4.3.1 Catchments and valley-bottom human pressure

4.3.1.1 Catchments land use

Two data sources were used to document land use changes over the last two centuries in the studied catchments. The first data corresponds to the état-major map (1832), whose georeferencing and digitisation of land use classes was carried out by INRAE (Favre et al., 2017). Historically, the état-major map was surveyed in $32 \times 20 \text{ km}$ zones by different surveyors (Dupouey et al., 2007). The choice of different land cover classes may therefore vary from one zone to another. For example, two land cover classes may be distinguished in one zone, while they are combined in the adjacent zone. In our study, the classes had to be simplified in order to have a consistent classification within the three catchments. Moreover, the studied catchment areas are covered by six different sheets of the état-major map, surveyed between 1830 and 1833 (Dupouey et al., 2007, <http://www.gip-ecofor.org/cartofora/index.php?sujet=cartes&ssmenu=source>). We use the average date of 1832 in our figures.

The second is the 'OCcupation du Sol à Grande Echelle' (OCSGE) land use map (2019) provided by the Grand Est region. This data was produced by photo-interpretation of colour and infrared aerial photographs. Its classification was simplified, in order to have the same classes as the ones retained for the état-major map. Sankey diagrams were computed using 'networkD3' R package (Allaire et al., 2017).

4.3.1.2 Production of a spatio-temporal database of hydraulic structures

Weirs The goal here is to reconstruct the spatio-temporal edification of weirs by producing an updated and spatialised database of every current and former weir present in

the main channel of each studied river. Two sources of data were used: the French national database of obstacles to flow (Référentiel des Obstacles à l'Ecoulement; ROE; <https://geo.data.gouv.fr/en/datasets/624470b851c2d7d1e2eb2aad149abb1cf9438e03>) and documentary archives (18-20th).

The ROE has been produced by the Office Français de la Biodiversité (OFB) since 2010 in partnership with local stakeholders. This regularly updated spatial database lists all flow barriers present in France. The obstacles are classified into four types: dams, river weirs, dykes, and bridge-induced obstacles. Each obstacle is assigned a unique identifier along with additional information, such as its current use, condition, and drop height (Léonard et al., 2014). The 2019 version of the ROE used in this study lists a total of 110,268 obstacles to flow in France.

The documentary archives used in this study are housed in the Departmental Archives of Haut-Rhin (Colmar, France) and Vosges (Epinal, France). In France, the classification of the archives in the departmental services is similar from one department to another. This means that they can be consulted in a relatively similar way from one department to another. The archives are always arranged by series. For reasons of reproducibility and comparison between the three studied catchments, we focused on a single series of archives, the S series 'Travaux Publics' (civil engineering), in which are kept the documents produced by the 'Service hydraulique des Ponts et Chaussées'. The latter was in charge of the development and monitoring of waterways during the 19th century. The S series contains all the documents related to the regularisation of existing or authorisation of new hydraulic structures (i.e., dikes, bridges, weirs) built on river channels (Mesmin et al., 2024). Each request is stored in a folder. The S series covers the period from the beginning of the 19th to the first part of the 20th century (~1940). The records in this series are generally grouped under the headings 'Factories and Mills' or 'Watercourses and Factories' and have the advantage of being arranged by river and municipality. Each folder contains, almost systematically, detailed plans for each of the works concerned, making it possible to locate them and sometimes giving useful information about several characteristics of riverbeds at a very local scale (Payrastre, 2005). Some folders may also be accompanied by a verification report, which attests the project was completed. These pieces of information help to reduce uncertainties about the building date of weirs.

In this study, the ROE was used to draw up a current inventory of the weirs on each of the studied rivers. The ROE was checked and cleaned using a 1 m LiDAR-derived DEM (RGE Alti), whose altimetric accuracy (0.2-0.7 m; IGN) enables us to detect weirs currently present in the riverbed. The position of weirs in the ROE was verified and adjusted if required. In this case, they were consistently relocated at the top of the waterfall. Any duplicate was deleted. Any potential weirs visible on the DEM but not present in the ROE were considered

natural, thus not included in the ROE. The ROE was then updated and completed following the prospecting of documentary archives, whose methodology is described below.

All folders in the S series pertaining to requests for hydraulic structures along the three studied rivers were retrieved from the Departmental Archives, amounting to approximately 17 meters of stacked paper documents (Fig. 4.5). All folders were reviewed, with particular attention given to those containing at least one plan. From these folders, the following information was extracted: the location of the site, the date of the plan, the municipality, the folder number, the intended purpose of the planned hydraulic structure, and the presence of a verification report. In QGIS, the plans were located using an ancient (mid-20th) orthorectified topographical map (SCAN50, IGN). We used place names, road and bridge locations, and the morphology of the channel depicted in the plans. Through this process, some of the weirs associated with the hydraulic structures were successfully relocated.

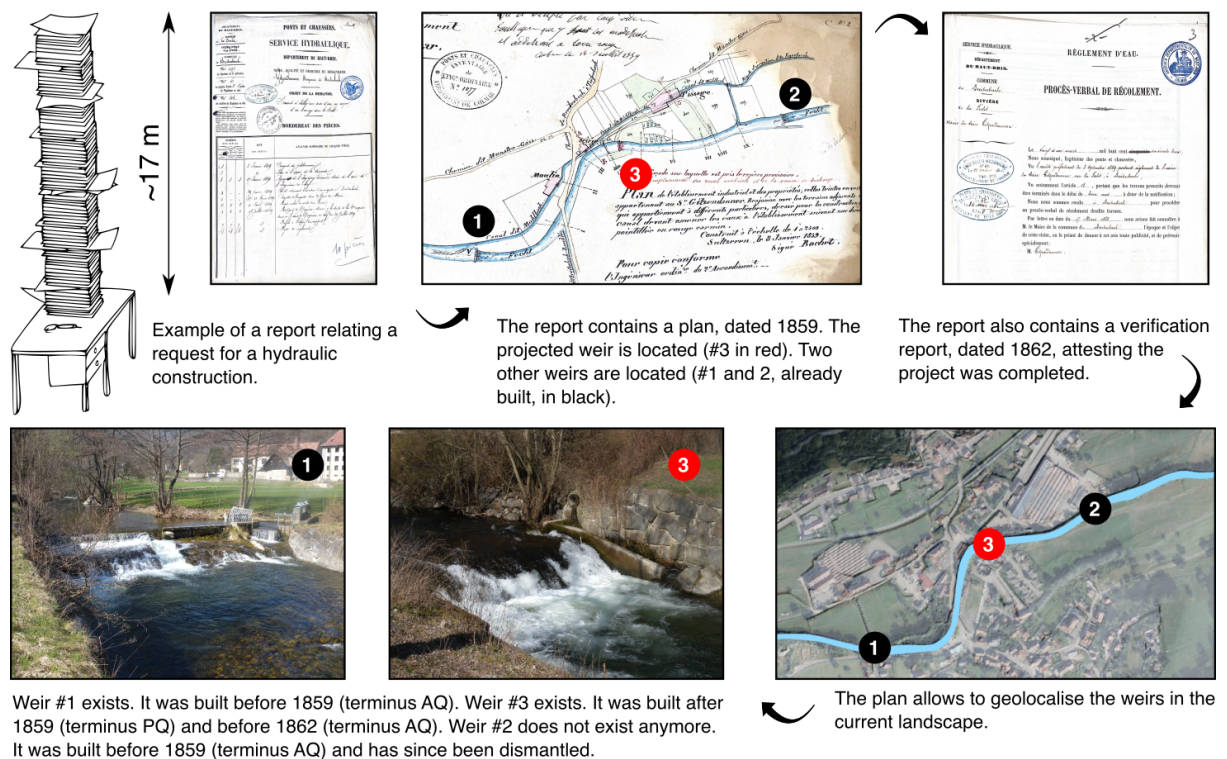


FIGURE 4.5: Workflow of the methodology used to attribute *termini ante* or *post-quem* to the weirs found in documentary archives.

The terminology '*terminus ante-quem*' (AQ) and '*terminus post-quem*' (PQ) has yet to be explicitly defined for the purpose of assigning construction periods to the hydraulic structures. A *terminus* AQ represents the latest possible date for the occurrence of an event or the existence of an item, while a *terminus* PQ denotes the earliest possible date. Associating a *terminus* AQ with a hydraulic structure thus implies that the structure was constructed before the *terminus*. Conversely, associating a *terminus* PQ with a hydraulic structure implies that the structure

was constructed after the *terminus*. Regarding the determination of construction periods for the hydraulic structures from the documentary archives, three potential scenarios exist.

1. A hydraulic structure appears on a plan as a construction project, accompanied by a verification report. The date of publication of the plan corresponds to a *terminus* PQ, and the date of the verification report corresponds to a *terminus* AQ (Mesmin et al., 2024). In this case, the period of construction of the weir is relatively well-constrained. If the associated weir does not exist today (invisible on DEM), the weir has been destroyed after the *terminus* AQ. It is also categorised as currently absent.
2. A hydraulic structure appears on a plan as a construction project but is not accompanied by a verification report. If the associated weir exists today (visible on DEM), the date of publication of the plan corresponds to a *terminus* PQ. If the associated weir does not exist today (invisible on DEM), it was never built or has been built then destroyed. No *terminus* can be assigned. The weir is not added to the database.
3. A hydraulic structure appears on a plan as already existing. If the associated weir still exists today (visible on DEM), the date of publication of the plan corresponds to an *terminus* AQ (weir #1, Fig. 4.5). If the associated weir does not exist today (invisible on DEM), the date of publication of the plan corresponds to an *terminus* AQ, but the weir has been destroyed and is also categorised as currently absent (weir #2, Fig. 4.5).

Bridges and dikes Bridges and dikes across and along the main channels were digitised from the 2018 orthophotograph and the DEM, respectively. A construction period was allocated to each bridge, determined by its presence or absence on the état-major map, and the 1951 (1956 for the Fêche) and 2018 orthophotographs. Bridges were thus categorised into one of the three following periods: (1) built before 1832, (2) built between 1832 and 1951, and (3) built after 1951. No construction periods could be assigned to the dikes.

4.3.2 Channel planform evolution, valley morphometry and longitudinal granulometry

In order to document the evolution of the channel pattern of the rivers since the 19th century, three planimetric data were used: the état-major map (mid-19th), a mid-20th-century orthophotograph, and a 21st-century orthophotograph, all provided by IGN. They have a resolution of 4, 0.5, and 0.2 m respectively. The production date retained for the état-major map is 1832 (see part 3.1.1). The mid-20th and 21st-century orthophotos were surveyed in 1951 (1956 for Fêche) and 2018, respectively.

The active channels of the three rivers were digitised using QGIS. In case of braided reaches, every low flow channel was digitised (Piégay et al., 2009). Digitisation was carried out from downstream to upstream, starting at the confluence and continuing until the active channel was no longer visible upstream on the mid-20th orthophoto. The état-major map represents water channels in the form of either surface or linear features. In this study, we only digitised water channels represented by surface features. The total extent of the hydrosystems was created using the 'delete holes' tool in QGIS. The active channels and the total extent of the hydrosystems were then divided into 500 m elementary reaches (ER), following transects generated orthogonally to the centrelines of the valley bottoms (e.g. David et al., 2016; Beck and Corbonnois, 2016). The number of channels per ER was reported, corresponding to the number of times the upstream transect crosses a channel. For each ER, the area of the active channels and the hydrosystems extent were computed. The average width of the active channel(s) was calculated by dividing their area by the ER length (i.e. 500 m) and by the number of channel(s). The width of the hydrosystem was calculated by dividing their area by the ER length (i.e. 500 m). A multi-threaded index was calculated, adapted from the braided index (Mosley, 1981). The multi-threaded index was determined by dividing the total length of active (or low flow, in case of braiding) channels in the floodplain by the section length.

The valley-bottoms of the three catchments were manually digitised, bounded by hillslopes, thus including any alluvial and/or fluvioglacial terraces. Valley-bottom mean width was computed at each ER, by dividing the valley-bottom area by the ER length (i.e. 500 m). The presence of glacial-related deposits in the valley-bottom was determined according to the BRGM geological map. Mean active channel slope per ER was derived from the 1 m-resolution DEM (RGE Alti, IGN) using QProf QGIS plugin (Alberti et al., 2016).

Longitudinal evolution of the surface riverbed granulometry was assessed *via* Wolman's (1954) pebble-count method, at approximately one site every 2 to 3 km along the studied rivers, in order to represent the main longitudinal surface grain size variations (total of 48 sampling sites, corresponding to a mean spacing of about 2.5 km). Measures were always done on riffles (or rapids or runs for the most upstream sites), in order to acquire comparable data. The sites were chosen to avoid areas heavily impacted by hydraulic structures, while documenting the potential geomorphological legacies as thoroughly as possible. We randomly selected 100 clasts at each site and measured their b-axis. Due to potential glacial remnants in certain studied areas, boulders (>256 mm) were found in the riverbed. Our sampling methodology includes these larger clasts. Although we could only measure their apparent b-axis, which provides a conservative estimate of their size, their presence affected the spatial distribution of bed material sizes. This, in turn, affected the applicability of the Wolman pebble-count method (Bunte and Abt, 2001). However, they allowed us to assess the relative abundance of glacial-related deposits within the riverbed. This aligns with our sub-objective of exploring the potential control of glacial legacies on fluvial morphodynamics.

4.4 Results

4.4.1 Human pressure

4.4.1.1 Spatio-temporal evolution of catchments land use

Figure 4.6 depicts the spatio-temporal evolution of land use in the three studied catchments, from the mid-19th century (1832) to nowadays (2019). Seven land use classes have been used. However, it is important to note that the 'crops and grasslands' and 'crops' classes only apply to the 19th and 21st centuries, respectively. This is due to the classification used on the état-major map, which does not distinguish between crops and grasslands. The Sankey diagrams facilitate visualisation of the spatial redistribution of land uses.

In the mid-19th century, the forest cover in France was at a historical minimum (Dupouey et al., 2007). At that time, the catchment areas of the Fecht, Moselotte, and Vologne rivers were 42%, 40%, and 32% forested, respectively. By the beginning of the 21st century, these proportions had increased to 57 %, 69 %, and 59 %, respectively. Currently, the Moselotte catchment is the most forested, with over two thirds of its surface area covered by forest. The Vologne catchment, which was the most forested in the mid-19th, experienced the strongest afforestation, with the forest area almost doubling in less than two centuries. In all three catchments, almost all of the forest area present in the 19th century remained forested today. The increase in forest area in all three cases is attributed to the afforestation of former grasslands and crops.

In the mid-19th century, the areas under grassland and/or crops corresponded to 48 %, 59 %, and 67% of the catchment areas of the Fecht, Moselotte, and Vologne, respectively. It is worth noting that cultivation was primarily located on the slopes during this period (Rochel, 2004; Keller, 2023). In contrast, today, the small proportion of lands still cultivated is mainly located in the valley bottoms. The proportions are 14% for the Fecht (including vineyards), less than 1% for the Moselotte, and 4% for the Vologne. The current cultivated areas are not a result of changes in land use, but rather remnants of previous crops. This also applies to the vineyards in the Fecht basin, which have remained in the same location and proportion.

Finally, the valley bottoms in all three catchments have undergone significant urbanisation. The proportions of artificialised areas have increased from 2%, <1%, and <1% in the 19th century to 11%, 10%, and 10% in the 21st century for the Fecht, Moselotte, and Vologne, respectively.

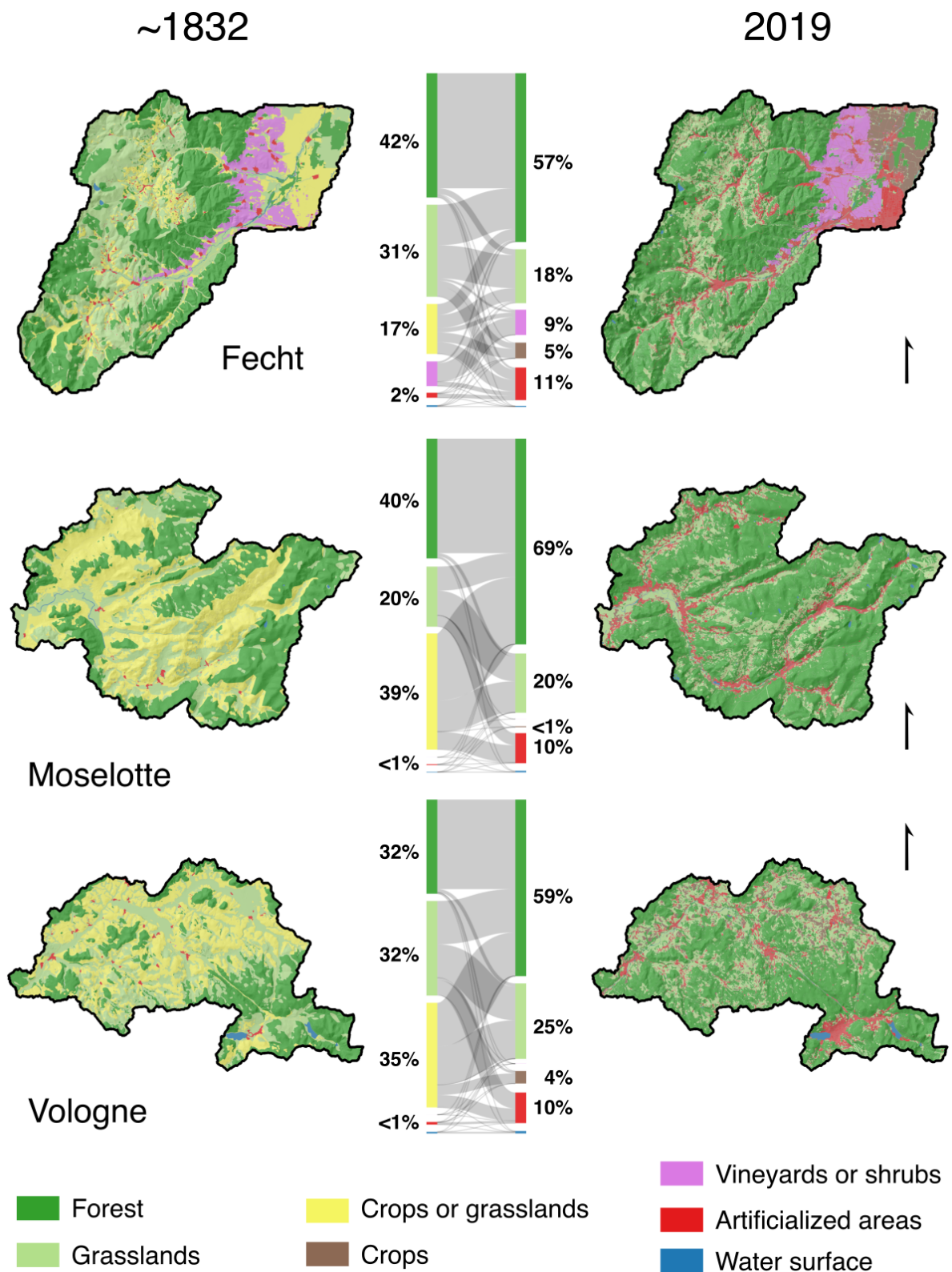


FIGURE 4.6: Land use evolution of the Fecht, Moselotte and Vologne catchments from the mid-19th to the 21st century. Sankey diagrams are associated to each pair of maps, to help visualize the spatial redistribution of land use classes.

4.4.2 Weirs updated database

A total of 225 folders containing at least one plan have been retrieved from departmental archives of Haut-Rhin (Colmar, France) and Vosges (Epinal, France). This survey is exhaustive and encompasses all folders (within the S series, see part 4.3.1.2) containing requests referring to the regularisation of existing or authorisation of new hydraulic structures along the entire length of the studied rivers. Figure 4.7 provides an overview of the temporal distribution of all these folders, by decades and by rivers. A total of 123, 57 and 45 folders were found for the Fecht, Moselotte and Vologne, respectively, with the decade presenting the highest number of folders being 1860-1870, with 36, 14 and 15 folders, respectively. The folders relating to the Fecht, Moselotte and Vologne range from 1780 to 1890, 1750 to 1900 and 1720 to 1960, respectively. Vologne and Fecht displays the largest (240 years) and the smallest (110 years) temporal distributions of folders, respectively.

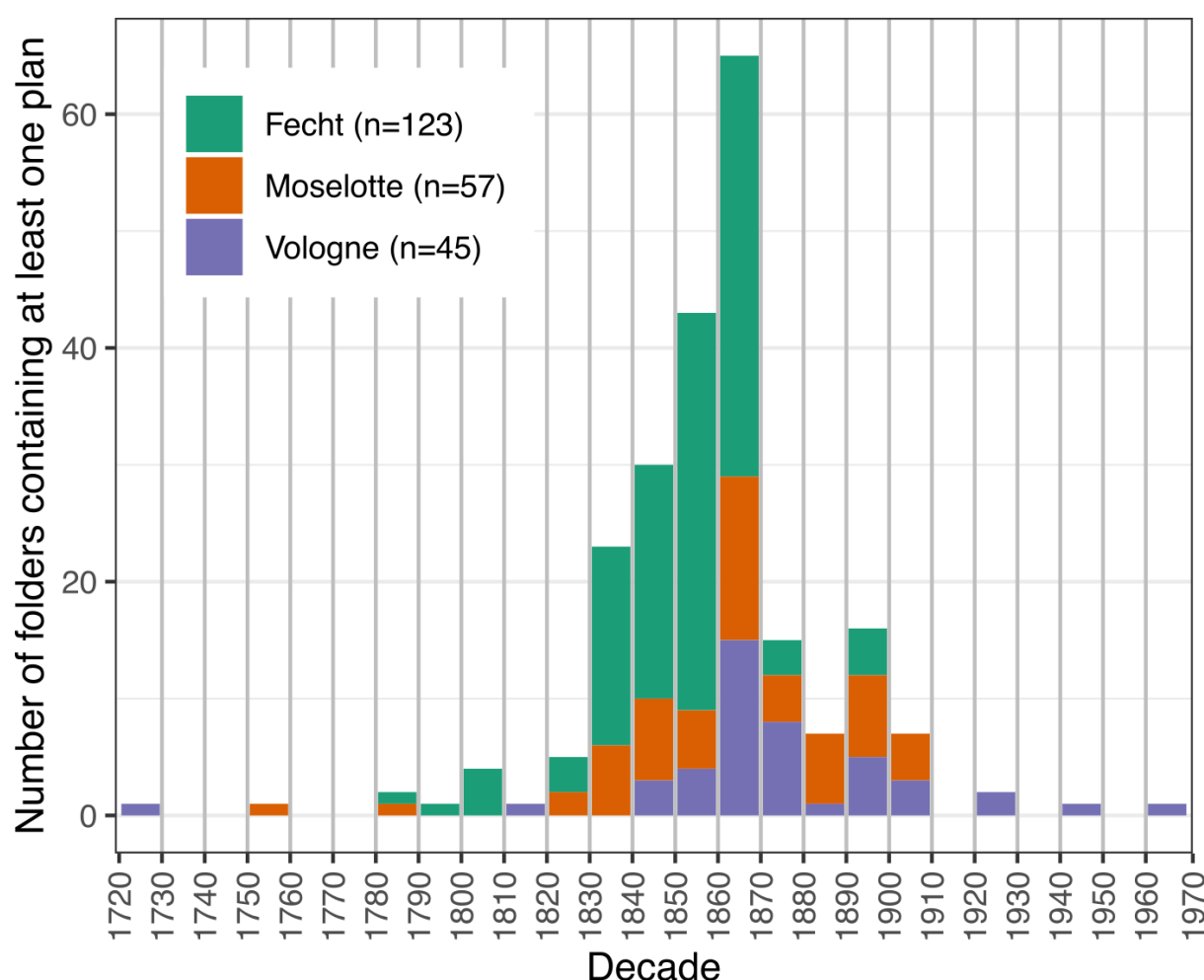


FIGURE 4.7: Temporal distribution of every folders containing at least one plan, found in documentary archives, for each studied river.

Some plans contained in these folders could be relocated within a GIS. It was possible to associate some of the weirs visible on these plans with weirs currently present in the studied

ivers (and therefore in the ROE database). Among the weirs currently present in the main channel of the Fecht, Moselotte and Vologne rivers, 10, 32 and 41% were found in the archive plans, respectively (Fig. 4.8A). Consequently, a *terminus* (AQ and/or PQ) has been assigned to them. Conversely, a total of 4, 13 and 21 weirs currently absent from the main channel of the Fecht, Moselotte and Vologne were found in the archive plans, respectively (Fig. 4.8B). Thus, a *terminus* (AQ and/or PQ) for the construction of these former weirs has been assigned to them. Figure 4.8C and 4.8D illustrate the temporal distribution of the *termini* AQ and PQ for each of the three rivers. The *termini* AQ for the Fecht, Moselotte and Vologne rivers span 19 years (1855-1874), 64 years (1836-1900) and 156 years (1818-1974), respectively. The *termini* PQ for the Fecht, Moselotte and Vologne rivers span 28 years (1836-1864), 70 years (1830-1900) and 113 years (1859-1972), respectively.

4.4.3 Channels longitudinal and lateral human pressure

Figure 4.9 synthesises the longitudinal distribution of all hydraulic structures (i.e. weirs, bridges and dikes) inventoried in this study and currently present along the Fecht, Moselotte and Vologne main channels and valley-bottoms.

The main channels of the Fecht, Moselotte, and Vologne rivers are crossed by a total of 125, 59, and 55 weirs, respectively. Weirs are relatively evenly distributed along the course of the Moselotte and Vologne rivers. However, the Fecht exhibits a more irregular distribution of weirs, with 94% of them located along only 63% of its length (ER 26-70).

The Fecht and Moselotte rivers have a similar number of bridges, with 42 on the Fecht and 45 on the Moselotte, while the Vologne has slightly fewer, with 33. Bridges are distributed relatively evenly along the Fecht and Vologne rivers. Meanwhile, the Moselotte river has a more irregular distribution, with almost 70% of its bridges located along only 40% of its length (ER 43-70).

The main channels of the Fecht, Moselotte, and Vologne rivers are surrounded by a total of 37 km, 17 km, and 21 km of dikes, respectively. The majority of dikes for all three rivers are located downstream. The downstream halves of the Fecht, Moselotte, and Vologne rivers represent 75%, 88%, and 62% of the total length of dikes, respectively.

4.4.4 Planform evolutionary trajectory

The fluvial planform evolutionary trajectory of the Fecht, Moselotte, and Vologne is depicted from 1832 to 2018 through three metrics: the active channel(s) mean width (Fig. 4.10), the hydrosystem mean width and the number of channels (Fig. 4.11), per 500 m-long ER.

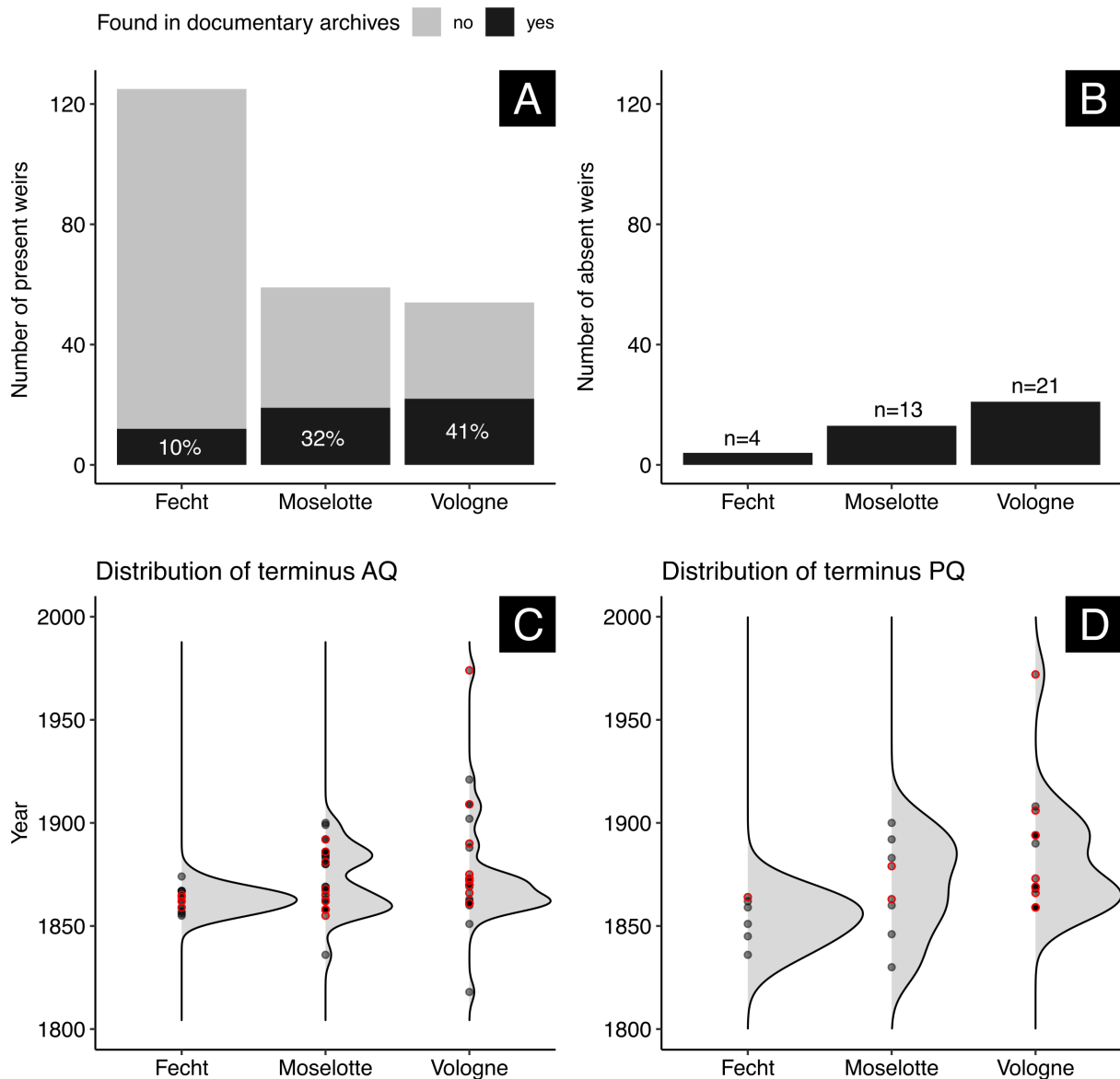


FIGURE 4.8: Contents of the weirs database produced in this study, which combines the French national database of obstacles to flow (ROE, 2019) with the data retrieved from documentary archives. (A) Number of weirs currently present in the main channel of each studied river, with the proportion of those found in the documentary archives. (B) Number of weirs found in documentary archives, which are not present anymore. For all weirs found in archives (currently absent in red), temporal distribution of the *terminus* (C) AQ and (D) PQ that have been assigned to them from the plans.

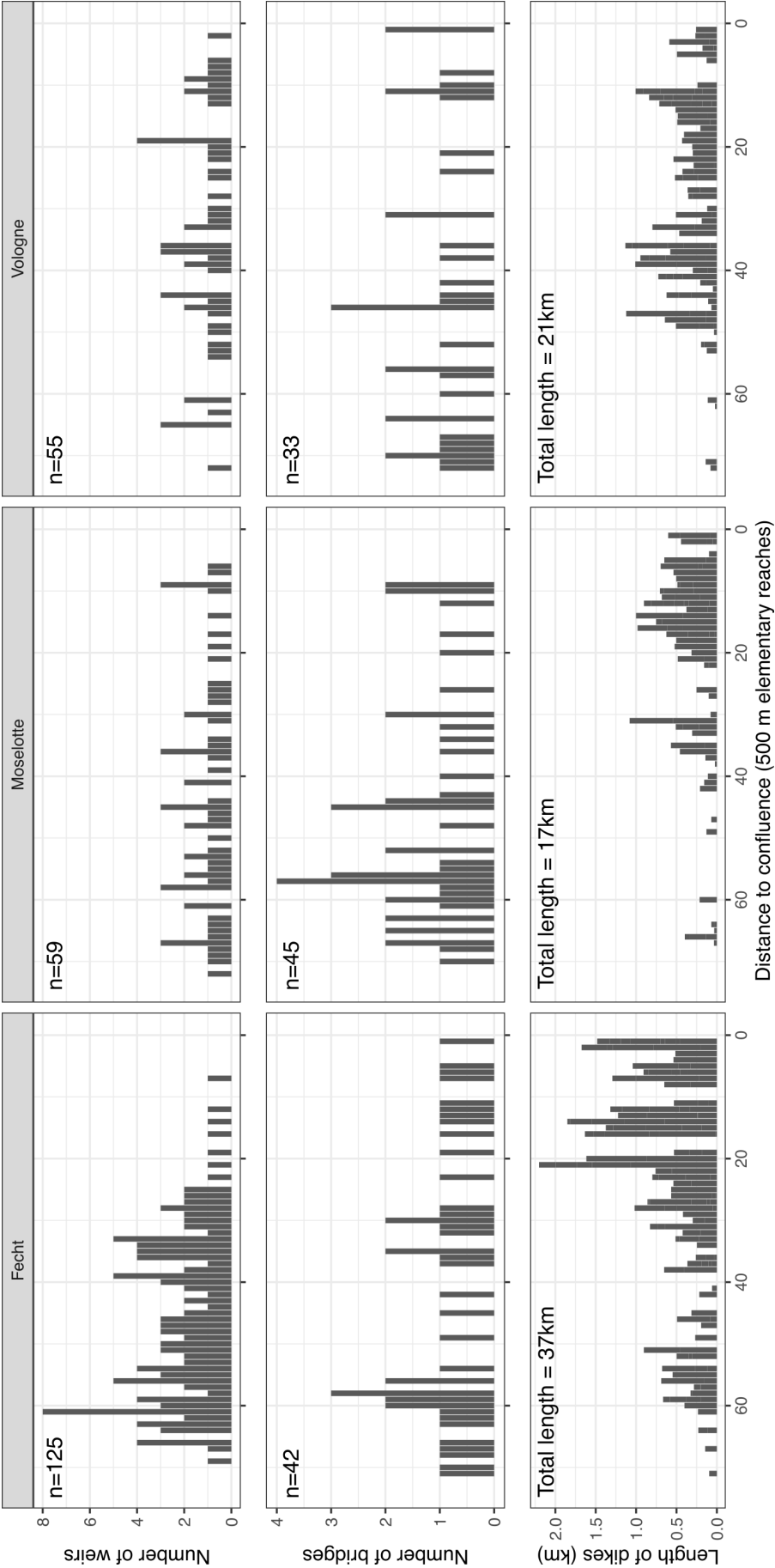


FIGURE 4.9: Longitudinal distribution of the number of weirs, bridges and length of dikes currently present along the main channel of the Fecht, Moselotte and Vologne rivers. Note that are not included the weirs identified in archives but currently absent.

4.4.4.1 Active channel(s) mean width

The average width of the active channel(s) of the Fecht decreased from 29.7 to 27.7 to 20.5 m between 1832, 1956 and 2018 (Fig. 4.10). The decrease in the average value of the active channel(s) width between 1832 and 1956 was relatively moderate (-2 m). However, there were strong decreases around ER #9 (-35 m), #16 (-32 m) and #33 (-83 m), and strong increases around ER #28 (+60 m) and #40 (+45 m). Between 1956 and 2018, there was a sharp decrease in the average width of the active channel(s) around ER #14 (-28 m), #28 (-70 m), #40 (-37 m) and #55 (-15 m). Elsewhere, the change in width does not appear to be significant.

The average width of the active channel(s) of the Moselotte decreased from 35.4 to 23.7 to 23.2 m between 1832, 1951 and 2018 (Fig. 4.10). It is important to recall that these values only refer to ER #1 to #46, i.e. the only ER that could be digitised on the état-major map (1832). Between 1832 and 1951, the average width of the active channel(s) in these ER decreased overall (-12 m), then did not change significantly between 1951 and 2018 (-0.5 m on average). On the other hand, the average width of the active channel(s) upstream of ER #46 seems to have decreased significantly over the same period (-3 m).

The average width of the active channel(s) of the Vologne decreased from 22.6 to 14.0, then increased a little to 15.0 m, between 1832, 1951 and 2018 (Fig. 4.10). Between 1832 and 1951, the average width of the active channel(s) along the entire length of the Vologne decreased by an average of about 9 m. Between 1951 and 2018, there was a slight increase in the average channel(s) width (+1 m), with local increases of up to 6 m (ER #40), 8 m (ER #28), 12 m (ER #16) and 15 m (ER #3).

In order to make a more detailed assessment of any changes in channel pattern and dynamics over the last two centuries, it is important to consider other metrics. For example, an average active channel width of 40 m could represent either a single channel of 40 m width or a system of four active channels of 10 m width each. Similarly, the metric used in Figure 4.10 does not take into account the distance separating multiple potential channels. Figure 4.11 addresses this issue by showing, for each ER, the total width of the hydrosystem associated with the number of active channels.

4.4.4.2 Hydrosystems mean width

The average width of the hydrosystem of the Fecht decreased from 189.0 to 70.8 to 55.2 m, while the average number of channel(s) decreased from 1.83 to 1.21 to 1.17 m, between 1832, 1956 and 2018, respectively (Fig. 4.11). Between 1832 and 1956, the average width of the hydrosystem fell by more than half (~120 m). Strong local decreases are observed around ER #21-27 (-1300 m), #46 (-110 m), #56 (-105 m) and #62 (-142 m). Two local increases are observed

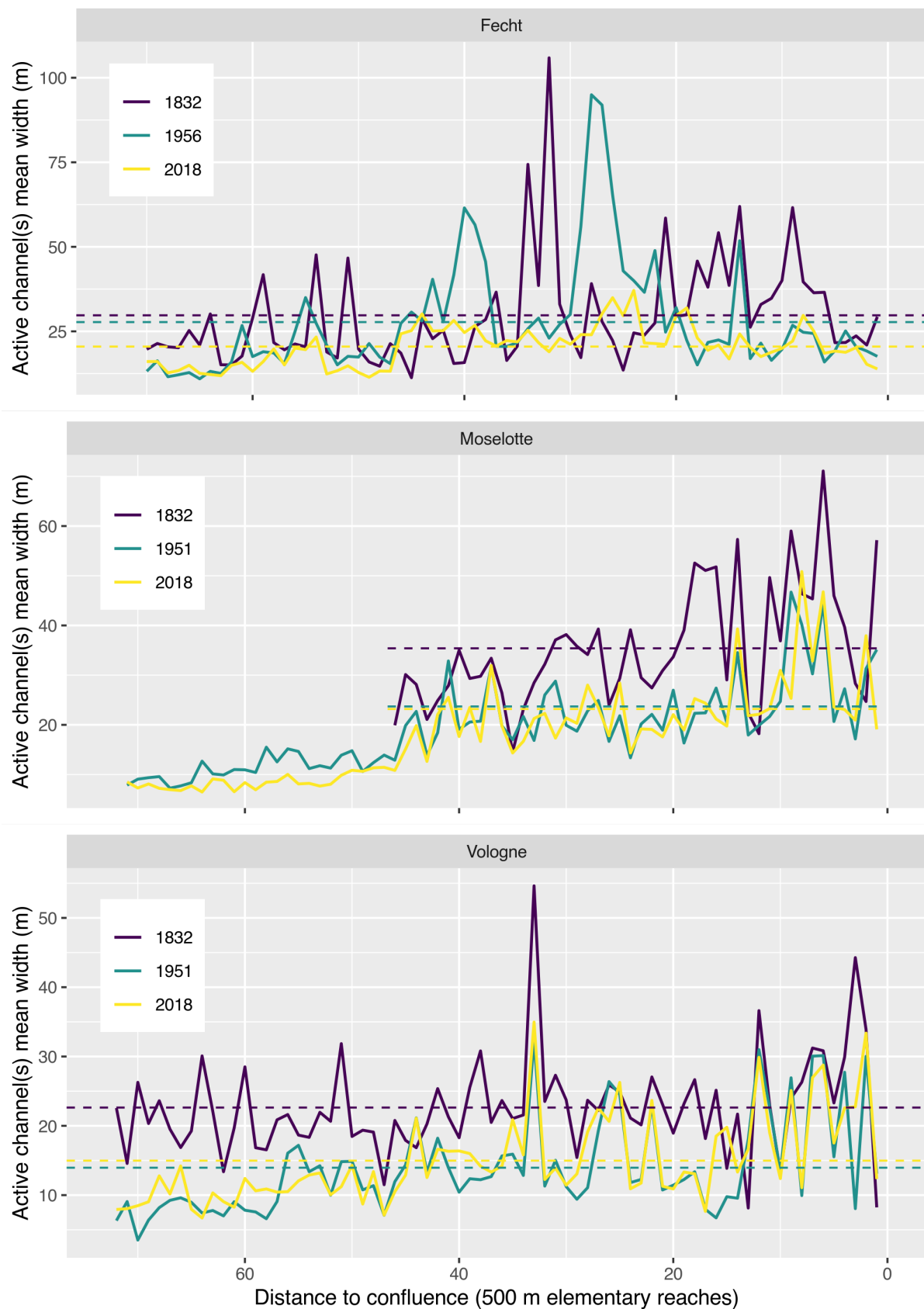


FIGURE 4.10: Spatio-temporal evolution of the active channel(s) mean width of the Fecht, Moselotte and Vologne rivers, from 1832 to 2018. Dashed horizontal lines represent the mean values for each date.

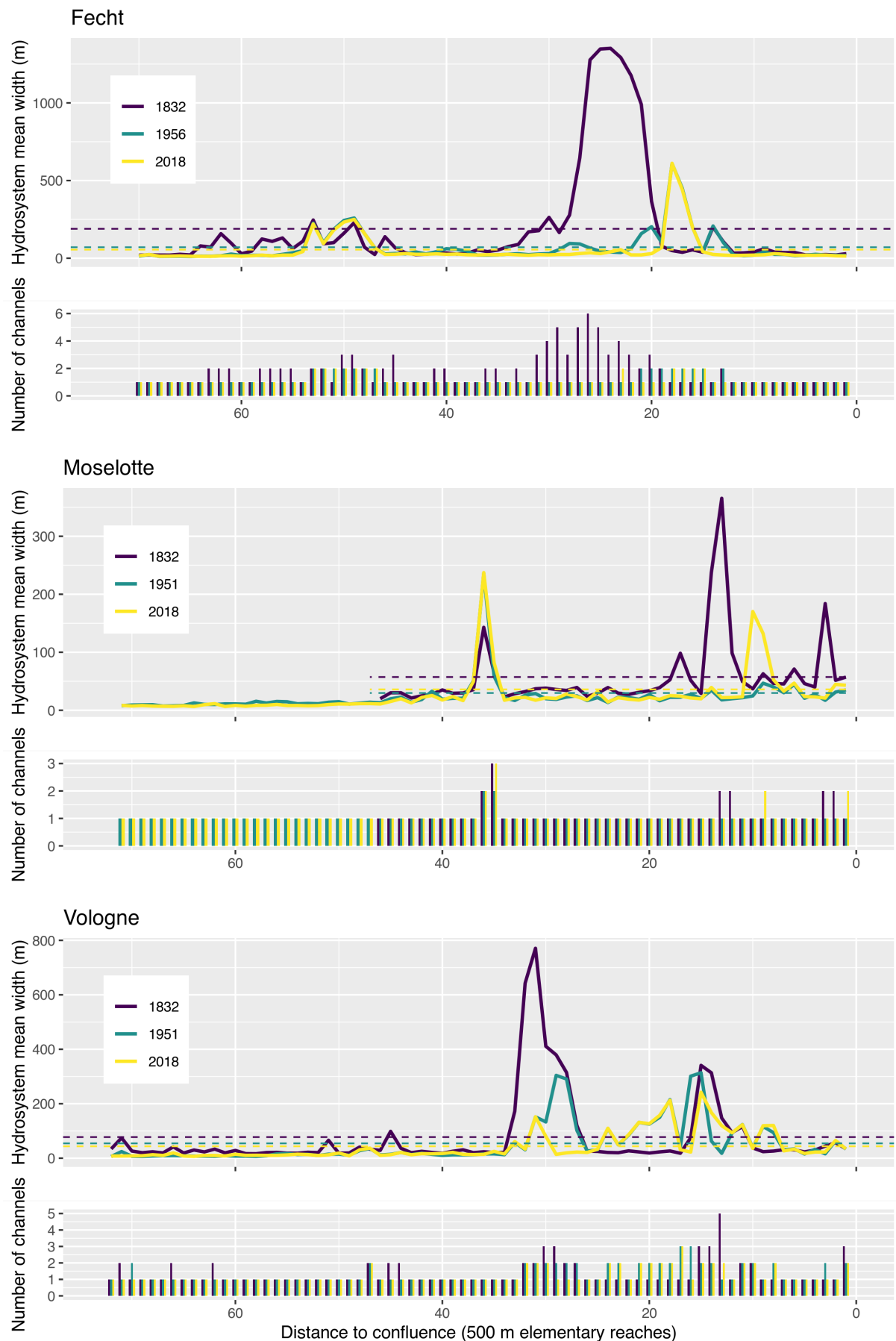


FIGURE 4.11: Spatio-temporal evolution of the hydrosystem mean width and the number of active channels of the Fecht, Moselotte and Vologne rivers, from 1832 to 2018. Dashed horizontal lines represent the mean values for each date.

around ER #18 (+560 m) and #51 (+86 m). Between 1956 and 2018, the smaller decrease in the average width of the hydrosystem (~ 16 m) is associated with a few localised width decreases, around ER #14 (-183 m), #20 (-173 m) and #28 (-71 m). Otherwise, no significant changes are observed.

The average width of the hydrosystem of the Moselotte decreased from 57.3 to 29.8 to 36.0 m between 1832, 1951 and 2018. These values only refer to sections 1 to 46, i.e. the only sections that could be digitised on the état-major map (1832). The average number of channel(s) decreased from 1.15 to 1.02 to 1.07 m between 1832, 1956 and 2018 (Fig. 4.11). Between 1832 and 1951, the average width of the hydrosystem decreased by almost half (-28 m). Strong decreases are observed around ER #3 (-167 m), #13 (-347 m) and #17 (-76 m), while an increase is observed around ER #36 (+93 m). Between 1951 and 2018, the average width increased by 20% (~ 6 m). Local increases are observed around ER #10 (+146 m) and #35 (+23 m). A slight decrease is observed upstream of ER #46, averaging around 3 m.

The average width of the hydrosystem of the Vologne decreased from 77.6 to 54.5 to 43.8 m, while the average number of channel(s) decreased from 1.36 to 1.35 to 1.26 m between 1832, 1956 and 2018, respectively (Fig. 4.11). Between 1832 and 1951, the average width of the hydrosystem decreased by around 30% (-23 m). A sharp decrease was observed around ER #31 (-620 m). Smaller, localised decreases are observed around ER #45 (-84 m), #51 (-51 m) and #71 (-50 m). Increases are measured around ER #9 (+92 m) and #18-24 (+100 m). Between 1951 and 2018, the average width decreased by a further 20% (-11 m). Two decreases are located around ER #16 (-279 m) and #29 (-290 m). Otherwise, no significant changes are observed.

The analysis can be refined by considering the number of channels per section. It is important to note that changes in the width of the hydrosystem, whether spatial or temporal, do not necessarily correspond to changes in the number of channels. Additionally, when the number of channels is equal to 1, the width of the hydrosystem (Fig. 4.11) will be the same as the average width of the active channels (Fig. 4.10). In this case, a change in hydrosystem width equals the change in active channel width. These details are used in the next section to deepen our analysis.

4.5 Discussion

The historical data used in this study are subject to different kinds of potential biases. Before attempting to interpret and compare the evolutionary trajectory of the three studied rivers in response to human pressure, it is essential to discuss the impact of these potential biases on the analysis of the results. The first part discusses the limits of using documentary archives and historical planimetric data, in general and in our specific study. Taking these limitations

into account, the second part discusses and compares the evolutionary trajectory of the three rivers, notably in their longitudinal dimension, in response to human pressure.

4.5.1 Data and methodologies: general limits and specific considerations

4.5.1.1 Documentary archives

We have thoroughly searched all the documents in the S series of the Haut-Rhin and Vosges Departmental Archives, which cover the entire length of the Fecht, Moselotte, and Vologne rivers. While most of the information relating to river engineering from the beginning of the 19th century to the end of the 20th is stored in this series (Payrastre et al., 2006; Jacob-Rousseau, 2015), our results can obviously not be considered exhaustive, neither temporally, nor spatially. Documentary archives are, by nature, fragmentary and are never produced with the aim of answering the questions that researchers (geographers or historians) ask themselves centuries later. In contrast, they are produced to fulfil particular requirements under specific circumstances (Furst, 2019). In our case, the documentary archives of the S series were generated primarily for any project related to river regularisation/channelisation or the authorisation of constructing new hydraulic structures, as weirs and dams (Payrastre, 2005). Some requests may also pertain to usage conflicts, grievances, or plans to repair structures (Jacob-Rousseau, 2015). In these archives, while the mapping of planned weirs seems obvious, that of weirs already built is less so. On the plan drawn at that time, the decision of whether to show all the weirs already present on the channel was left to the engineer in charge of mapping. Additionally, paper documents are susceptible to damage, destruction, or loss. Furthermore, while the management of Departmental Archives in France is standardised, the situation in eastern France, specifically Alsace/Lorraine, is unique. Due to the frequent shifting of the Franco-German border from the early 19th century to the present day (see study area), the management, production, and conservation of archives may have been consequently affected. This is all the more relevant as the territories of Alsace (that includes Fecht catchment) and Lorraine did not always change occupiers simultaneously. This was the case in 1871 after the Franco-Prussian War when Alsace came under German occupation, while the Vosges department, which includes the Moselotte and Vologne catchments, remained in French territory (Roth, 2014). Beyond a potential dichotomy in the management/production of documentary archives, the history of the studied valleys may also be responsible for some variability in industrial dynamics. It is known, for example, that the annexation of Alsace in 1871 led to an exodus of Alsatian industrial, who relocated to the Vosges department to pursue their activities (Muller, 1995; Parmentier, 2019; Vaillot, 2020). The decrease in population density (Fig. 4.4) and in the number of plans produced between the 1860s and 1870s (Fig.

4.7) may thus be attributed to the Franco-Prussian war and its impact on archive production and/or the industrial dynamics of the valleys.

The number of folders containing at least one plan found for the Fecht, Moselotte and Vologne rivers is 123, 57 and 45 (Fig. 4.7), respectively. In contrast, the number of weirs that could be located, including those currently present (Fig. 4.8A) and absent (Fig. 4.8B), amounted to 16, 32 and 43, respectively. The dichotomy between the number of folders found for each river versus the number of weirs that were relocated raises two potential major biases. Firstly, it should be noted that not all of the plans found could be geolocated. This may be due to the poor quality of the mapping, or its poor state of conservation. We also assume that changes in channel planform may make it difficult, if not impossible, to relocate certain plans. Secondly, some of the plans may have been redundant, when large number of folders were relating to the same channel reach. Also, the plans could represent several weirs, or just one, or none at all. For weirs that can be relocated from archive plans, a *terminus* (AQ and/or PQ) could be assigned to them, mostly AQ. It corresponds to the observation of a weir mapped as already existing on an archive plan (weir #1, Fig. 4.5). Of the weirs found in archives and relocated on the Fecht, Moselotte, and Vologne rivers, 63%, 75%, and 57% were assigned a *terminus* AQ. However, the assignment of a *terminus* does not specify exactly when the weir was built. For weirs whose both *termini* AQ and PQ were assigned, the median difference between the two *termini* is 2 years. Therefore, we can assume that a *terminus* PQ provides the actual date of construction of the weir within a range of two years. It is important to note that in some cases, only a *terminus* PQ could be assigned. This occurs when a weir is projected on a plan, is currently present in the channel but the file does not contain any verification report. In this case, we consider that the two-year rule mentioned above can be applied. However, if only a *terminus* AQ is assigned, this simplification is not possible. Indeed, if the weir is represented on the plan at a certain date, it may have been present for decades or even centuries (Furst, 2022). It should be noted that the potential pressure on the channel was exerted at the latest from the *terminus* AQ, but most likely before. For the period of construction of the weir, we thus argue that the *terminus* PQ is more informative than the AQ, although the latter are more numerous. A comparison of the temporal distributions of *terminus* PQ (Fig. 4.8D) suggests a difference in the temporal dynamics of industrial activities in the Fecht, Moselotte, and Vologne valleys. The mean (interquartile range) *terminus* PQ for the Fecht, Moselotte, and Vologne are 1853 (15), 1869 (29), and 1882 (32), respectively. Assuming that the distribution of *terminus* PQ is representative of river development periods during the 19th century, the Fecht was developed over a relatively short period of about 15 years, concentrated in the 1850s, whereas the Moselotte and Vologne rivers were developed later, in the late 19th century, over a period of about 30 years.

Finally, we must also consider (i) the absence of temporal data concerning the majority of weirs currently present on rivers and (ii) the current absence of weirs that were actually built

(i.e. found in archives). It should be remembered that the studied archives mainly cover the 19th-20th centuries. In case (i), two distinct assumptions can be made: - These weirs were present before the 19th century and were not mentioned (mapped) in the archives. - These weirs were all built after the 19th-20th century.

The proportion of weirs meeting case (i) is 90%, 68%, and 59% for the Fecht, Moselotte, and Vologne rivers, respectively (Fig. 4.8A). The probability that all these weirs were built after the 19th century seems low. Especially as it is known that hydraulic structures have been present in the VM long before the industrialisation period. For instance, in 1773, a survey commissioned by the Intendance d'Alsace identified approximately 100 medieval mills in the Fecht catchment (Furst, 2022), all of which being used for various purposes such as grain, oil, sawmills, and metallurgy. On the western side of the VM, Parmentier (2019) lists around 30 mills during the 15th century. Therefore, we can consider the first assumption, which suggests that the weirs are older than the 19th century and were not mentioned in the archives. As previously stated, the inclusion and mapping of weirs is significantly influenced by archive production conditions. However, concerning the Fecht, Wohlgemuth (1986) has demonstrated that the political and administrative chaos caused by the French Revolution (1789) resulted in the near-total destruction of hydraulic structures dating from before the 19th century. Consequently, it is probable that a significant proportion of the weirs not documented in the archives, but currently present, were constructed after the 19th or even during the 20th century. The number of weirs meeting case (ii) is 4, 13 and 21 for the Fecht, Moselotte and Vologne, respectively (Fig. 4.8B). This information is important, as the lowering of a weir, whether gradual or sudden, natural or human-caused, may have an impact on the fluvial system. The questions are: when, why and how were they destroyed? Unfortunately, the initial use of these weirs (factory or irrigation) doesn't provide any information. In fact, for each river, among the weirs that are missing, there are about as many whose initial use was agricultural irrigation as factory irrigation. The discrepancy in the number of weirs identified as having been destroyed post-19th century raises questions about the cause of their destruction. If the lowering was natural, we might expect that the number of weirs destroyed should be related with the respective energy of each river. However, it doesn't seem to be the case. In fact, the Vologne has the greatest number of weirs destroyed (21), while the Moselotte has the highest Q_2 (112 m³/s) and channel gradient (11.7‰; Table 4.1) for a lower number of destroyed weirs (13). Let's assume that the majority of absent weirs are related to voluntary human-made weirs lowering from the beginning of the 21st century. The significant difference in the number of weirs currently destroyed could then be explained by different management strategies. In our case, two river management syndicates share the three catchments. The Fecht is managed by one syndicate (Rivières Haute Alsace), while the Moselotte and Vologne are managed by another (Syndicat Mixte Moselle Amont). We can therefore hypothesise that the large number of weirs (21) destroyed in the Vologne, whose energy is relatively moderate,

is due to voluntary lowering undertaken by the syndicate. Conversely, the relatively small number of weirs (4) identified as destroyed on the Fecche may be indicative of an absence of willingness on the part of the syndicate to remove them, and/or the presence of relatively important human constraints that make impossible or difficult some weir removals.

4.5.1.2 Diachronic planform analysis

In this study, we aimed at reconstructing the evolutionary trajectory of the three studied rivers using three planimetric data, two of which being historical planimetric data, i.e. the état-major map and the mid-20th century orthophotograph. These data offer the advantage of being available for the entirety of France in an orthorectified format, providing a historical snapshot of river planforms in the mid-19th and mid-20th centuries. They have been widely used for this purpose in the literature for several decades (Gurnell et al., 1994; O'Connor et al., 2003; Hooke and Yorke, 2010; Morais et al., 2016; Arnaud et al., 2019; Jautzy et al., 2022). However, it should be recalled that these data are affected by a non-negligible geometric error (Herrault and Sheeren, 2022). More recently, in fluvial contexts specifically, a few authors have shown that it is crucial to quantify and consider the uncertainty associated with historical planimetric data (Lea and Legleiter, 2016; Donovan and Belmont, 2019). This is especially important when measured surface changes are used to determine the significance of fluvial planform changes. Two reproducible methods have been proposed to characterise the uncertainty associated with surficial changes in a river context (Leonard et al., 2020; Jautzy et al., 2020). However, these methods are only applicable to channel planform changes in the case of a single channel rivers. In the present study, the variety of channel pattern displayed by the three rivers, both temporally and spatially, unfortunately makes this type of approach impossible. This is why we have chosen to investigate the evolutionary trajectory of rivers using various metrics (Fig. 4.10,4.11,4.12), including metrics that are not affected by geometric error. A more generic approach proposed by Jautzy et al. (2024a) might allow for the association of uncertainties with the measured widths of channels (Fig. 4.10) or hydrosystems (Fig. 4.11). In their study, Jautzy et al. (2024a) measured the uncertainties associated with land use polygons digitised from a mid-20th century orthophotograph and from the état-major map, i.e. the same historical planimetric data used in our study. The polygon areas ranged from 0.0027-0.029 km² and 0.011-0.446 km², respectively. They found that the total uncertainties associated with polygons digitised from 20th century orthophotography were less than 5%, and that polygons digitised from the état-major map exhibit total uncertainties ranging from 4-40%. In the present study, the area of the digitised polygons ranges from 0.003-0.68 km², indicating that they are likely affected by similar uncertainties as those measured by Jautzy et al. (2024a), i.e. 4-40% for the one digitised from the état-major map. Therefore, we must consider these uncertainties when interpreting our results, especially regarding the width of active

channels in 1832. Indeed, Jautzy et al. (2020) demonstrated a negative correlation between surficial uncertainty and the amplitude of measured change. In contrast, our measurements of hydrosystem(s) width (Fig. 4.11), which exhibit stronger changes than channel(s) width, are less susceptible to be affected by these uncertainties. Moreover, the other metrics used in this study, i.e. number of channels (Fig. 4.11) and multi-threaded index (Fig. 4.12B), are not surface-based. They are therefore not affected by the geometric error associated with the data.

Concerning the analysis of land use changes based on the état-major map, its geometric error may obviously have an impact on our results. However, we argue that given the large scale of the analysis (whole catchment), and the mean area of the digitised polygons ($\sim 1 \text{ km}^2$), uncertainties are expected to have only a minor impact on our measurements. Finally, it should be remembered that the non-distinction of certain land-use classes used on the état-major map (i.e. 'grasslands' and 'grasslands or crops'), unfortunately prevented us from accurately studying the spatial redistribution of these two classes.

4.5.2 Impact of human pressure on the evolutionary trajectory of river channels

This section discusses the evolutionary trajectory of each of the three rivers, taking into account various human controlling factors, at different spatial scales, and specific morphological characteristics of each fluvial system. Without trying to be exhaustive, we describe in detail some notable fluvial changes for each of the three rivers, which illustrate the various cause-and-effect relationships that can be highlighted from the data produced in this study. Evolutionary trajectories are also compared between rivers to potentially help to disentangle the respective share of each controlling factor, whether human or not. The evolutionary trajectory of the channel patterns of the three studied rivers is depicted in Figure 4.12B, over 200 years and along 500 m ER, considering (i) human impact, quantified by the number of weirs and bridges, as well as the length of dikes (Fig. 4.12A), and (ii) the morpho-structural controls of the valley bottoms, including valley bottom width and presence of (fluvio-)glacial deposits. Channel slope and bed surface grain size, a semi-dependent and dependent variables, respectively, are also integrated in order develop a complete morphodynamic diagnosis (Fig. 4.12C).

4.5.2.1 Evolutionary trajectories of the Fecht, Moselotte and Vologne

Over the last two centuries, the Fecht channel pattern experienced a significant fluvial metamorphosis, almost generalised along its entire length (Fig. 4.12B). The average value of the multi-threaded index declined from 2.2 in 1832 to 1.4 in 1956, and further to 1.3 by 2018. This

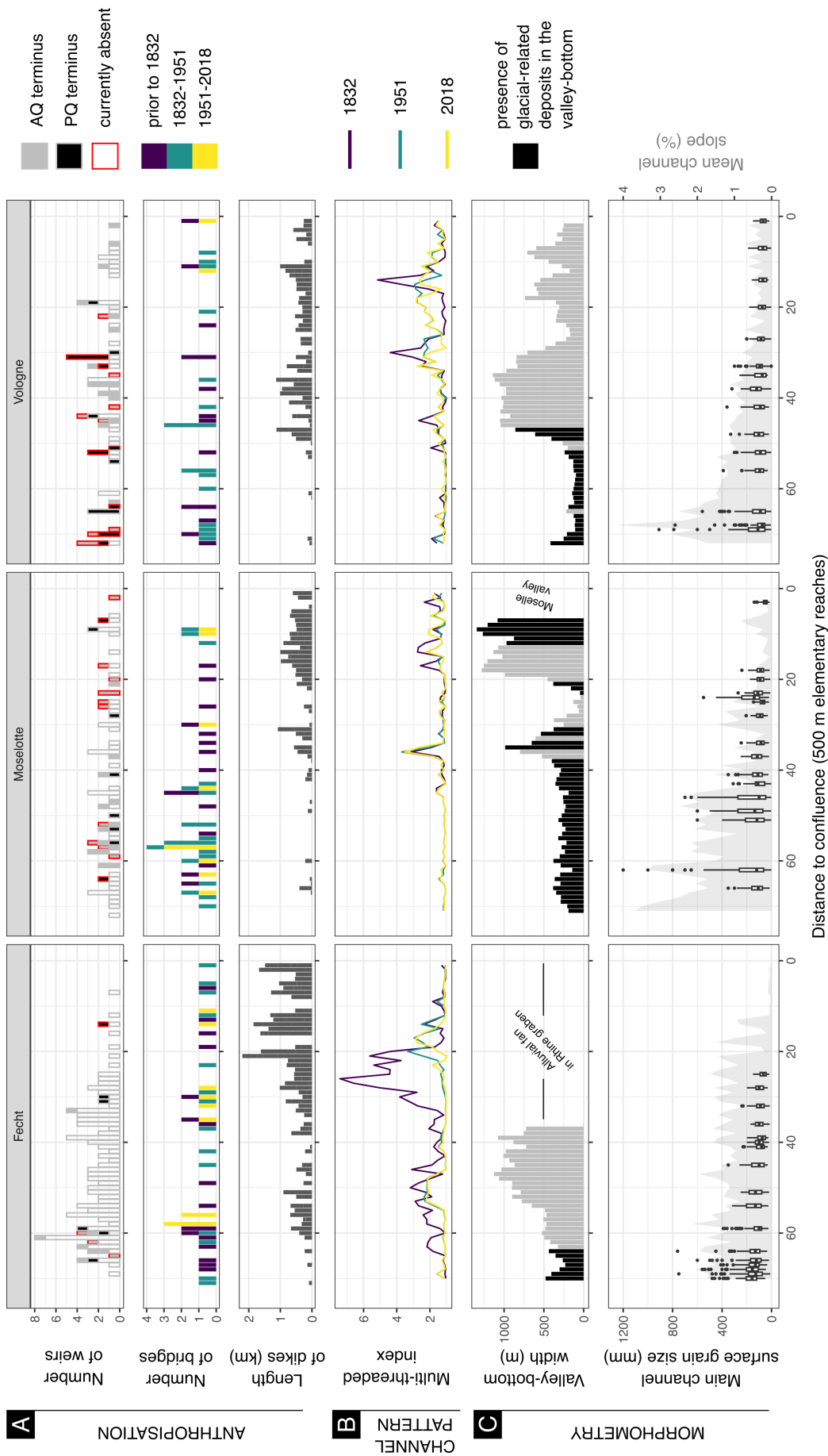


Figure 4.12: Spatio-temporal geomorphological diagnosis of the Fecht, Moselotte and Vologne rivers over the last two centuries. Longitudinal variability of (A) anthropisation, (B) channel pattern evolution and (C) valley-bottom width and channel slope and surface grain size.

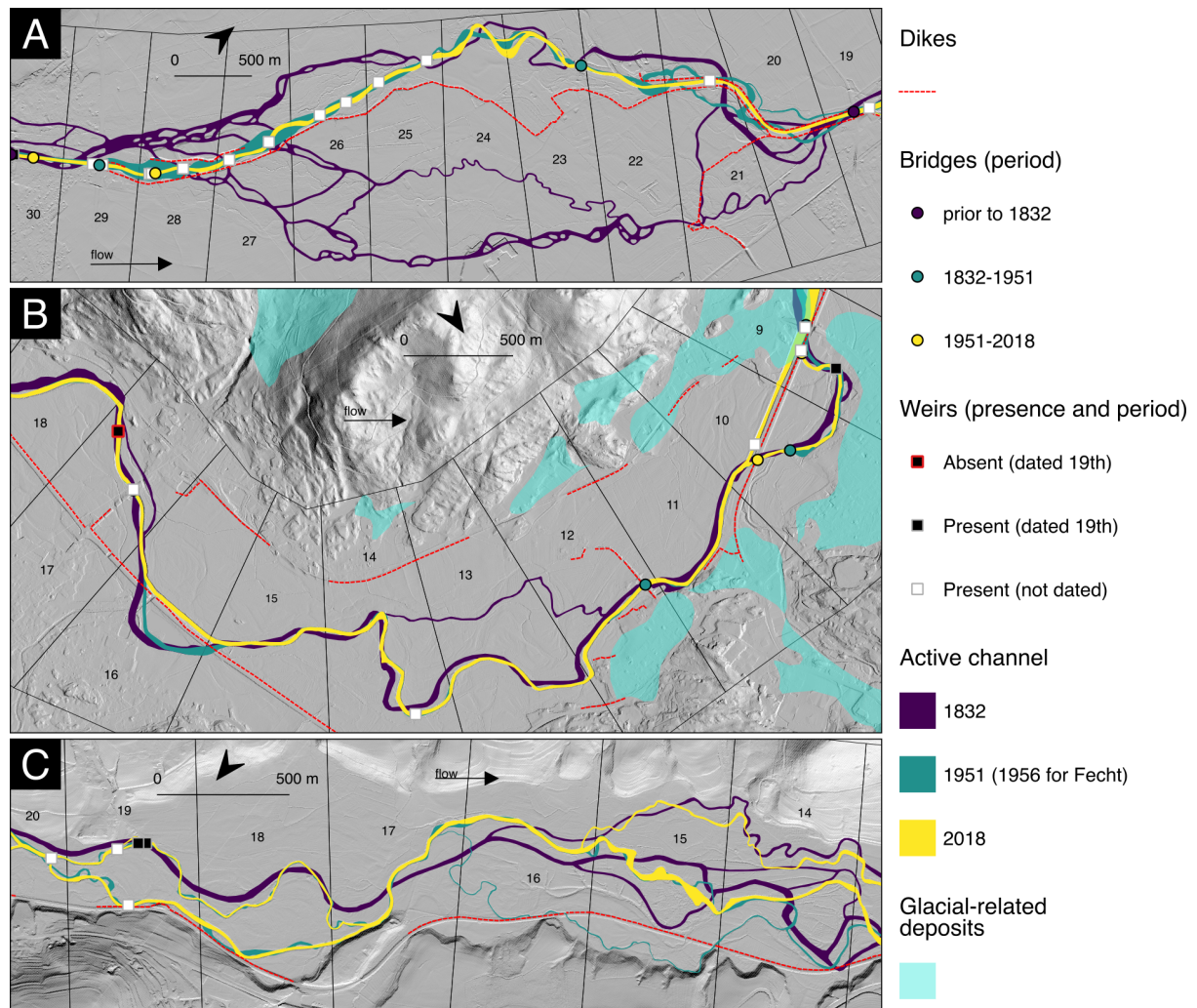


FIGURE 4.13: Planimetric extracts of the channel pattern evolution of the (A) Fecht, (B) Moselotte and (C) Vologne rivers, from 1832 to 2018, associated with hydraulic structures.

decrease in the average value actually masks much more significant localised changes. For example, the section located between ER #20 and #30 experienced a much greater decrease in the average value of the multi-threaded index, from 4.1 to 2.2 to 1.3, from 1832 to 1956 to 2018, respectively. In this sector, in less than a century, the Fecht evolved from a braided (ER #28 to #30) and anabranching (ER #20 to #27) channel pattern to a single channel (Fig. 4.13A). Taking into account both human and morpho-structural control factors, this localised fluvial metamorphosis provides some insights into the causes of both the former fluvial dynamics (i.e. 1832) and its evolution to nowadays. First, the former high fluvial dynamics, characterised by a large width of the hydrosystem (> 1000 m; Fig. 4.10) and a large number of channels (up to 6 channels; Fig. 4.10), was favoured by specific morpho-structural conditions. Indeed, this sector is located at the outlet of the massif, on a large alluvial fan with no lateral constraints by hillslopes, whose channel slope value ($\sim 1.5\%$) is identical to that inside the massif (Fig. 4.12C). Moreover, the relatively low river bed surface grain size (b-axis < 100 mm) and the absence of glacial legacies may favour a sustained lateral fluvial dynamic.

Second, the evolution of the channel pattern can be directly related to human actions. Indeed, this sector has a relatively large total length of dikes (~ 1 km/ER; Fig. 4.12A), the location of which (Fig. 4.13A) allows us to conclude without any doubt that they are responsible for the simplification of the channel pattern. This highly probable cause-and-effect relationship also suggests that these dikes have been erected between 1832 and 1951.

Over the last two centuries, no significant change has been measured in the channel pattern of the Moselotte, from ER #71 to #18 (Fig. 4.12B). The average value of the multi-threaded index only slightly decreased, from 1.5 in 1832 to 1.3 in 1951, and remained stable at 1.3 by 2018. The slight decrease in the average value of the multi-threaded index measured over the 19th century is explained by an overall decrease in river lateral dynamics, from ER #17 to #1, for which the average value of the index fell from 2.2 to 1.3 between 1832 and 1951. This sector is located in the area where the width of the valley-bottom is the largest (>1000 m; ER #17 to #7; Fig. 4.12C) as well as in the outlet into the Moselle alluvial plain (ER #6 to #1). It corresponds to the only sector of the Moselotte course where the river is unconfined, with the exception of a short widening of the valley-bottom between ER #37 and #35 (Fig. 4.12C). Combined with a relatively low channel slope ($<0.2\%$), the river had sufficient divagation space in the 19th century to allow for developing wide meanders. In this sector, the decrease in the value of the multi-threaded index around ER #17, #13 and #3 is related to the abandonment of a secondary channel. For ER #17, this may be due to the destruction of a weir (Fig. 4.13B), that was probably keeping in water the secondary channel. However, the cause of the weir's removal is unknown. Between 1951 and 2018, we also note a slight increase in the value of the multi-threaded index around ER #10, which is human caused. In fact, a meander cut-off was provoked by the erection of a lateral dike, while the former channel was kept in water (Fig. 4.13B).

Concerning the evolution of the Vologne channel pattern over the last two centuries, its trajectory appears less linear than for the Fecht and Moselotte (Fig. 4.12B). The average value of the multi-threaded index remained relatively constant, ranging from 1.5 in 1832 to 1.6 in 1951, and then returning to 1.5 by 2018. But this apparent stability actually masks some notable local changes. In the 19th century, there were two significant decreases in the multi-threaded index, at ER #30 and around ER #14. These reaches are located in sectors where the width of the valley floor (500-1000 m) and the relatively low channel slope ($<0.5\%$) allowed the river meander widely and developing a multi-threaded pattern (Fig. 4.12C). They correspond to anabranching sectors, where the channel positions changed and their number decreased between 1832 and 1951 (Fig. 4.13C). As no human action could be identified from our data, we can conclude that these anabranching channels evolved naturally. Conversely, also in the 19th century, a significant increase in the multi-threaded index is observed between ER #22 and #18. This is most likely related to the reconnection of a secondary channel caused by the

construction of a weir (ER #20; Fig. 4.13C), which has not been found in the archives, but whose construction date can thus be deduced (i.e. between 1832 and 1951).

4.5.2.2 Comparison of the evolutionary trajectories

The comparison of the evolutionary trajectory of the three rivers, in relation to the specific human pressure they experienced and their specific geomorphic context, allows more general findings. Over the last two centuries, the highest fluvial dynamics and consequently the greatest channel pattern variety, were observed during the 19th century for both the Fecht and Moselotte, whereas no clear trend has been identified for the Vologne. The river with the strongest (weakest) fluvial dynamics at the beginning of the 19th century is the Fecht (Moselotte), while fluvial dynamic of the Vologne river has remained relatively constant over the last two centuries. We have shown that this difference is clearly controlled by the distinct geomorphic context of each catchment, i.e. large valley-bottom to totally unconfined for the Fecht, and mostly confined for the Moselotte. Concerning the Fecht, we can consider that the river was almost completely free of any hydraulic structures at the beginning of the 19th century. In fact, Wohlgemuth (1986) has shown that the political and administrative chaos caused by the French Revolution (1789) resulted in the pre-19th century hydraulic structures being abandoned and almost completely destroyed, returning the Fecht to what he describes in his work as a 'wild state'. However, we consider that this pre-industrial state cannot be considered as a 'natural' state (Brown et al., 2018). Indeed, it should be remembered that our study began at the end of the Little Ice Age (LIA). Characterised by transitional hydrological conditions between a flood-dominated regime in the 19th to a drought-dominated regime in the 20th century (Bravard, 1989; Delile et al., 2016; Jantzi et al., 2017), it has been documented for the Rhine and some Vosgian tributaries, including the Fecht (Himmelsbach et al., 2015) and the Moselle (Bonnefont and Carcaud, 1997). Taking into account the impact of the LIA transition is all the more important in the context of developing strategies for restoring river systems, particularly in the light of future hydro-climatic projections. In addition, we have shown that the vegetation cover of the catchment areas has changed considerably (Fig. 4.6), with increases in forest cover ranging from +36% (Fecht) to +84% (Vologne). It is therefore very likely that the former sediment supply in the 19th century was much higher than today. The impact of the combination of the end of the LIA and the historical minimum forest cover on sediment supply, and consequently on river fluvial dynamics, has been well documented, particularly in the Alps (Bravard and Petit, 1997; Liébault et al., 2005), but also in and around the Vosges Massif (Bonnefont and Carcaud, 1997; Jautzy et al., 2021).

Regarding the anthropisation of the three rivers and their valley bottoms, our study demonstrates significant and various human actions across all three rivers over the studied period (19th-20th). However, their temporal trajectories exhibit slight variations (Fig. 4.8C-D). While

the pressure exerted by the construction of weirs on the Fecht river most likely peaked in the mid-19th, that exerted on the Moselotte and Vologne rivers occurred a bit later and extended over a longer period, continuing during the early 20th century. Our results also highlights a significant decreasing gradient of anthropisation among the Fecht, Moselotte, and Vologne valleys (Fig. 4.12A). Concerning the potential relationship between the degree of anthropisation and fluvial adjustments, some of the measured changes in the channel pattern of each river could have been clearly attributed to human influence. The most striking illustration documented in this study is the fluvial metamorphosis experienced by the Fecht (ER #30 to #20; Fig. 4.13A). On the other hand, some fluvial adjustments might be simply controlled by the specific geomorphic context of the valley-bottoms, as they couldn't be clearly related to any human driver. Although out of scope in this study, the impact of weirs on longitudinal hydrosedimentary connectivity should be assessed (e.g. Peeters et al., 2020). Finally, in order to further our interpretations, it would be important to consider also the hydro-climatic characteristics of the catchment, including any significant floods that might have occurred in recent centuries. This would offer insights into several important aspects, such as the channel pattern observed in the mid-19th century, the evolutionary trajectory to the present day, and the dynamics of anthropisation affecting the channels and valley bottoms over time.

4.5.3 Conclusions and perspectives

In this study, we aimed at assessing the extent to which the anthropisation of catchments, valley bottoms and channels of the Fecht, Moselotte and Vologne has affected their channels morphodynamics over the last two centuries. These three adjacent catchments are located in the southern part of the Vosges Massif, which is the most densely populated French massif. As a secondary goal, we tested the value of using documentary archives produced by the 'Ponts et Chaussées' administration to assign construction periods to the numerous weirs currently present in the river channels. The main findings of our study are as follows. A very marked decreasing gradient of pressure exerted by the construction of hydraulic structures has been demonstrated between the Fecht, Moselotte and Vologne rivers. Although the valley bottoms and channels of each river were significantly anthropised during the 19th century, the use of documentary archives made it possible to specify different temporal trajectories between the Fecht (eastern slope) and the Moselotte/Vologne (western slope). The first peak of human pressure in the 19th century appears to have been slightly earlier and shorter than for the Moselotte and Vologne. This difference is probably due to the singular history of the region, where the Franco-German border frequently shifted. Another major advantage of using archive documents is the possibility of identifying hydraulic structures that have been destroyed.

The evolutionary trajectory of rivers has shown a generalised simplification of the channel pattern of the Fecht, whose impressive fluvial metamorphosis has been measured in less than a century on its downstream section. The trajectories of the Moselotte and Vologne, on the other hand, are less impressive. Indeed, the Moselotte is highly confined in its valley bottom and the Vologne appears to have maintained a relatively high dynamic over the last two centuries. The human impact on the trajectory of rivers has been widely demonstrated, in particular a spatial correlation between the presence of dikes and the simplification of the channel pattern. On the other hand, certain fluvial adjustments could not be explained by human pressure. The study of the cause-and-effect relationships between human pressure and fluvial adjustments therefore illustrates the importance of taking into account the geomorphological context in order to understand past and current fluvial dynamics. This is particularly important in the context of managing and restoring channels and valley bottoms. To this end, we suggest that the pre-industrial state should not be considered as a 'reference state' to be targeted. Indeed, it has been shown that the hydro-climatic and vegetation cover context in the beginning of the 19th was very different from the current one. Given that the studied rivers can be considered in morphologically-novel configurations (Downs and Piégay, 2019), we recommend taking into account of the factors currently controlling their dynamics in order to better plan and anticipate the results of restoration actions (removal of weirs, embankment, dikes). We propose a number of research avenues to improve our understanding of river dynamics and the impact of human pressure. Firstly, modelling the temporal evolution of longitudinal hydro-sedimentary connectivity, taking into account changes in land use in the catchment areas (Jautzy et al., 2021), would make it possible to characterise the actual hydrogeomorphological sensitivity of the river to changes in land use. Then, to go further in assessing the influence of human and morpho-structural controlling factors on the evolutionary trajectory of the three rivers, a multivariate statistical analysis could be applied, as in David et al. (2016) and Arfeuillère et al. (2022). Finally, the longitudinal assessment of the specific stream power of the rivers would make it possible to prioritise reaches to restore.

Chapitre 5

Dynamique morpho-sédimentaire des rivières vosgiennes : synthèse, perspectives et recommandations

Ce dernier chapitre vise à synthétiser les principaux apports scientifiques de la présente thèse et à formuler des perspectives à la fois thématiques et méthodologiques, pour les chapitres 2, 3 et 4. Ensuite, des premières pistes de réflexions ainsi que des recommandations de gestion des rivières vosgiennes seront présentées. Elles s'appuieront sur les résultats obtenus dans les chapitres 2, 3 et 4 ainsi que sur des résultats préliminaires supplémentaires ayant été acquis dans le cadre de ce travail de thèse.

5.1 Facteurs de contrôle de la dénudation et du déséquilibre cosmogénique du massif des Vosges

L'étude du Chapitre 2 a permis de mettre en évidence, à l'échelle du massif des Vosges, que 41 % des bassins versants vosgiens (9/22) sont à l'état de déséquilibre cosmogénique (i.e. ratio $^{26}\text{Al}/^{10}\text{Be} < 6.7$). Ce résultat est d'autant plus important que la localisation de ces bassins versants témoigne d'une dichotomie Nord-Sud très marquée en terme de (dés)équilibre cosmogénique du massif (Fig. 5.1). Concernant les bassins versants à l'équilibre, 70 % d'entre eux (9/13) sont situés dans la partie nord du massif, dans la zone n'ayant jamais été englacée. Concernant les bassins versants en déséquilibre, 78 % d'entre eux (7/9) sont situés dans la partie sud du massif, dans la zone ayant été anciennement englacée. Notre étude a ainsi permis de démontrer une relation spatiale entre l'englacement passé du massif et le déséquilibre cosmogénique des bassins versants.

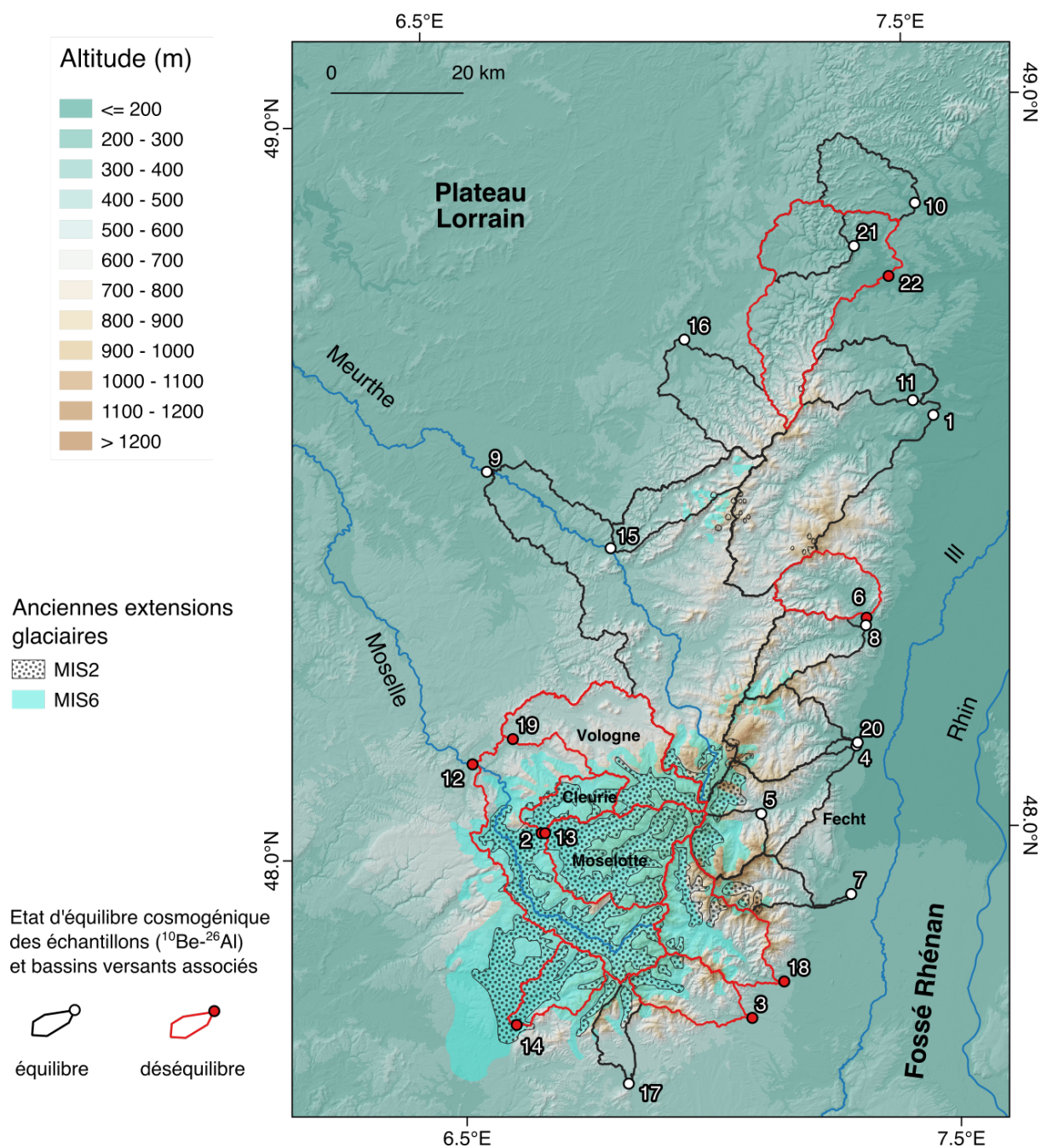


FIGURE 5.1: Dichotomie Nord-Sud en terme d'équilibre cosmogénique des bassins versants vosgiens. Mise en évidence de la relation spatiale avec l'englacement du massif.

Les taux de dénudation dérivés du ^{10}Be dans les bassins versants à l'équilibre sont compris entre 34 ± 1 (bassin #11) et 66 ± 2 (bassin #7) mm/ka (Fig. 5.2A). Ces valeurs sont comprises dans une gamme de valeurs comparable à celle des taux de dénudation mesurés dans d'autres massifs de moyenne montagne en Europe Occidentale, comme le massif de Bohême (Schaller et al., 2001), le Massif Central (Olivetti et al., 2016) et la Forêt-Noire (Meyer et al., 2010). Les ratios $^{26}\text{Al}/^{10}\text{Be}$ mesurés dans les sédiments prélevés dans le lit des rivières drainant les bassins versants en déséquilibre sont compris entre 1.4 (bassin #2) et 5.7 (bassins #6 et 22), ce qui correspond à des durées apparentes d'enfouissement comprises entre 3.4 et 0.5 Ma, respectivement (Fig. 5.2B).

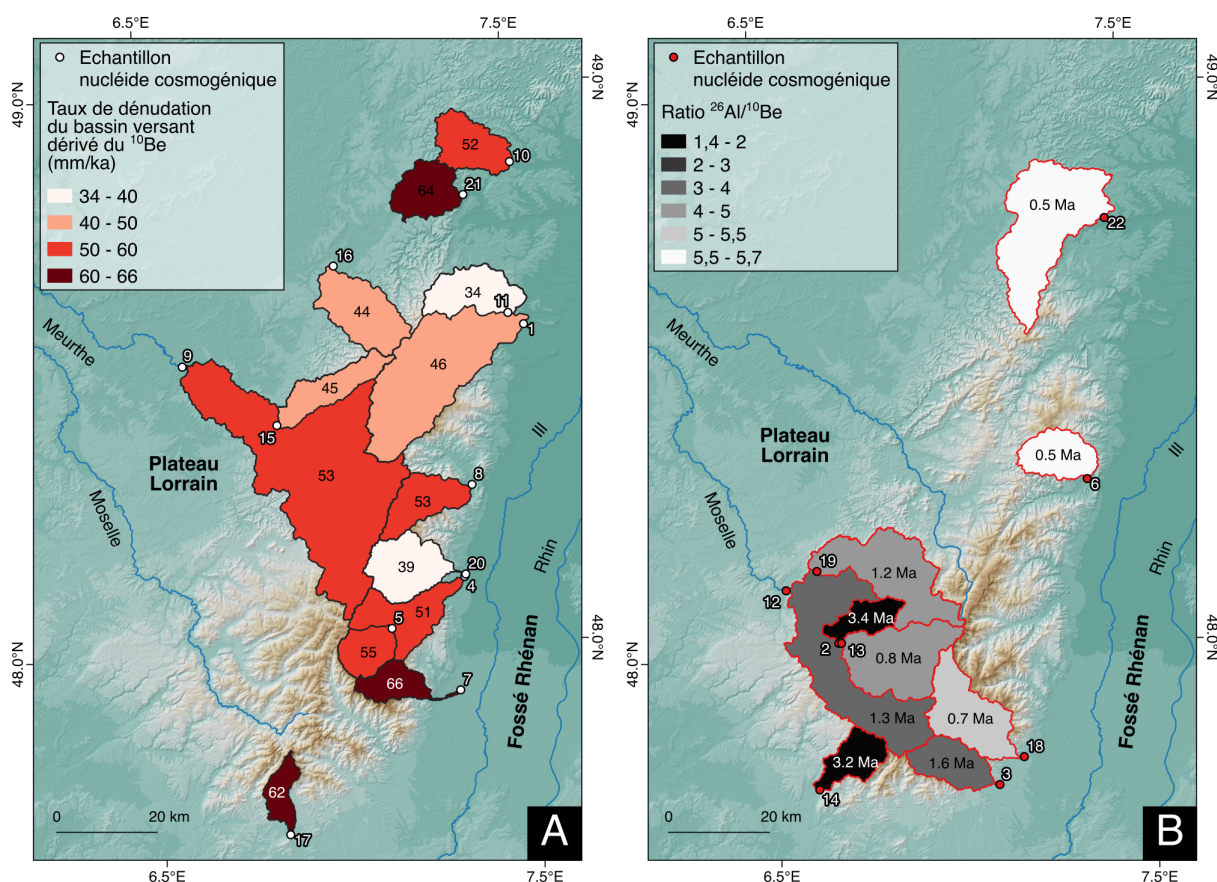


FIGURE 5.2: (A) Distribution spatiale des taux de dénudation dérivés du ^{10}Be parmi les bassins versants à l'équilibre. (B) Distribution spatiale des ratios $^{26}\text{Al}/^{10}\text{Be}$ parmi les bassins versants en déséquilibre. L'âge indiqué dans les bassins versants correspond à la durée apparente d'enfouissement des sédiments échantillonnés, en millions d'années, considérant un scénario d'enfouissement simple.

Concernant les facteurs de contrôle de la dénudation dans les bassins versants à l'équilibre, aucune relation significative n'a été constatée entre les principaux facteurs de contrôle couramment utilisés dans la littérature (altitude, pente, k_{sn} , précipitations) et les taux de dénudation mesurés (Annexe A; Fig. A.2). Cependant, nous avons suggéré un fort contrôle de la lithologie sur la dénudation des bassins versants à l'équilibre (voir partie 2.5.1.2). Premièrement, en se focalisant spécifiquement sur les bassins versants à lithologie dominante gréseuse d'une part, et plutonique d'autre part (Fig. 5.3A), nous observons en effet des taux

de dénudation moyens très similaires (48 ± 10 et 45 ± 9 mm/ka), malgré des valeurs moyennes de l'indice de pente du chenal (k_{sn}) significativement plus élevées pour les bassins versants plutoniques (60 ± 8 m^{0.9}) que pour les bassins versants gréseux (17 ± 7 m^{0.9}). Cela suggère très clairement une plus forte érodibilité du grès comparativement aux roches plutoniques. Deuxièmement, comme suggéré par Olen et al. (2016), lorsque la lithologie semble exercer un contrôle significatif sur la dénudation, il est probable d'observer des relations entre les facteurs de contrôle climatiques ou morphométriques et la dénudation, au sein d'un sous-échantillon de bassins versants dont la lithologie dominante est similaire. Or dans notre cas, parmi les 13 bassins versants à l'équilibre, 5 bassins versants sont à lithologie dominante gréseuse (Fig. 2.3). Ce nombre s'élève à 6, si l'on y ajoute le bassin versant #22 dont le faible degré de déséquilibre peut être considéré comme relevant de biais méthodologiques (Wittmann et al., 2020, Fig. 2.5). Contrairement à la corrélation positive généralement observée dans la littérature entre k_{sn} et la dénudation (e.g. Cyr et al., 2014), nous observons une corrélation significative négative au sein des bassins versants vosgiens gréseux (Fig. 5.3B). Plus l'indice de pente est élevé, moins la dénudation est forte. Cela suggère un découplage potentiel entre les sources (interfluves et parties sommitales du plateau) et les flux sédimentaires (versants et réseau hydrographique), qui serait alors fonction du degré d'incision. Ces deux observations témoignent indiscutablement d'une évolution singulière des paysages gréseux vosgiens (Buntsandstein). Elles ont donné lieu au dépôt d'un projet de recherche EUCOR - Seed Money ('Binding the Bunt'), porté par Gilles Rixhon, qui vise à caractériser et quantifier l'évolution quaternaire des paysages gréseux au sein du massif des Vosges et de la Forêt-Noire, ainsi que dans la Forêt du Palatinat (Nord du massif des Vosges) et l'Odenwald (Nord de la Forêt-Noire), afin d'approfondir l'analyse des deux observations effectuées dans les Vosges. Elle combinerait des mesures de concentrations en ¹⁰Be et ²⁶Al dans les sédiments de rivières (taux de dénudation) et dans la roche mère (durée d'exposition), sur les interfluves et parties sommitales du plateau gréseux. La durée d'exposition en surface des différents compartiments des bassins versants gréseux pourrait également être mesurée *via* des mesures de luminescence (Brill et al., 2021).

Concernant les facteurs de contrôle du déséquilibre cosmogénique, mesuré au sud du massif dans la zone anciennement englacée, nous avons mis en évidence une corrélation significative négative entre d'une part le ratio ²⁶Al/¹⁰Be et d'autre part l'extension glaciaire des deux dernières glaciations (MIS2-MIS6) ainsi que la proportion de dépôts (fluvio-)glaciaires présents dans les bassins versants (Fig. 2.8). Ces résultats impliquent nécessairement une histoire d'exposition complexe de la partie englacée du massif des Vosges, pouvant être expliquée par des héritages cosmogéniques dans le socle rocheux et/ou des épisodes d'enfouissement/stockage/remobilisation des sédiments dans les fonds de vallées (Wittmann et al., 2020).

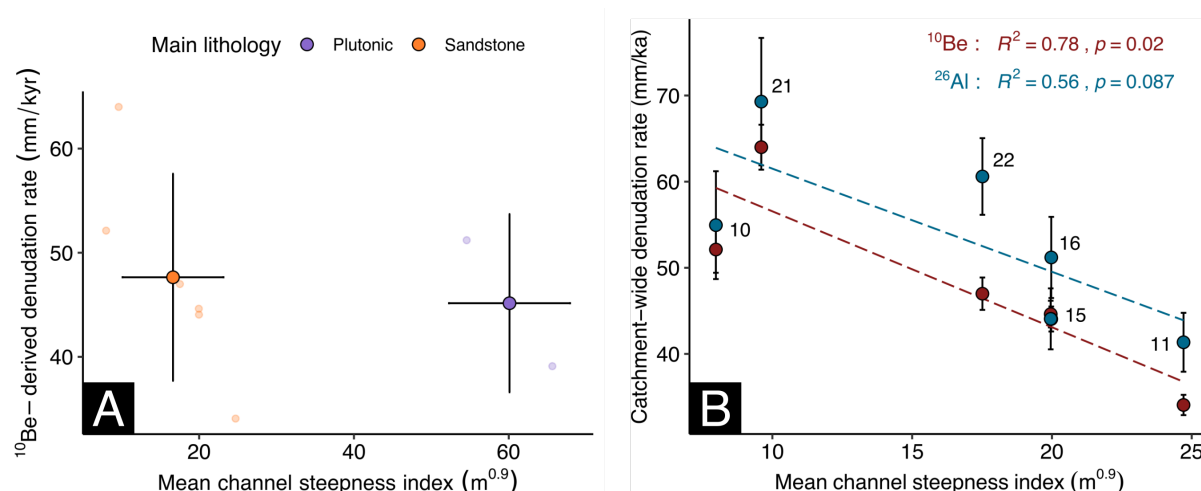


FIGURE 5.3: (A) Comparaison de la valeur moyenne ($\pm$ l'écart-type) du taux de dénudation en fonction de l'indice de pente du chenal (k_{sn}), parmi les bassins versants à l'équilibre, à lithologie dominante gréseuse et plutonique. (B) Corrélation négative significative entre le taux de dénudation et l'indice de pente du chenal (k_{sn}), parmi les bassins versants à l'équilibre à lithologie dominante gréseuse. Le bassin versant #22 a été ajouté au jeu de données, considérant que son faible degré de déséquilibre relève de biais méthodologiques (Wittmann et al., 2020).

En ce qui concerne l'impact potentiel du masquage de la roche mère par les anciennes glaciations, il conviendrait (i) d'améliorer la reconstitution de leur étendue spatiale respective, (ii) d'évaluer les paléo-épaisseurs de glace et (iii) d'établir une chronologie robuste de leurs empreintes respectives. Ce dernier travail a été effectué récemment par Valenti et al. (2024) mais de manière restreinte à quelques petits cirques, en tête du bassin de la Plaine (#15, Fig. 2.1E), aux alentours du Donon (Fig. 2.2B, photo b). Les auteurs ont pu confirmer que dans cette zone de transition entre la zone non englacée et la zone englacée, des petits glaciers de cirques étaient présents au DMG, sans dépasser 45 m d'épaisseur. La reconstitution des étendues glaciaires devrait également être améliorée en cartographiant précisément la totalité des erratiques et dépôts morainiques, qui sont abondants dans le secteur. Ensuite, des mesures de concentration en ^{10}Be et en ^{26}Al du substratum rocheux poli par les glaciers devraient également être effectuées pour évaluer l'efficacité de l'érosion glaciaire et vérifier l'hypothèse d'anciens glaciers à base froide. En ce qui concerne les modalités de transport, d'enfouissement et de remobilisation des sédiments dans les fonds de vallées, des mesures de concentration en ^{10}Be et en ^{26}Al devraient être effectuées le long des cours d'eau dans les bassins versants orientés vers l'est et vers l'ouest. La datation des dépôts des fonds de vallées et la réalisation de bilans sédimentaires des vallées vosgiennes apporteraient également des éléments clés. Enfin, dans les deux cas (masquage glaciaire et/ou enfouissement des dépôts), des modélisations numériques (e.g. Corbett et al., 2016; Knudsen and Egholm, 2018) pourraient permettre de reconstituer des scénarios d'exposition/enfouissement, en mesure d'expliquer les ratios $^{26}\text{Al}/^{10}\text{Be}$ particulièrement bas mesurés dans le massif des Vosges.

Ces perspectives de recherche devraient donc permettre d'expliquer comment les échantillons de sable prélevés dans le lit des rivières vosgiennes dont les bassins ont été englacés, peuvent présenter des durées apparentes d'enfouissement comprises entre 0.5 et 3.4 Ma (Fig. 5.2B). Il convient également de rappeler que de telles valeurs impliquent nécessairement que les cours d'eau aient été en mesure de remobiliser des dépôts ayant été enfouis pendant plusieurs centaines de milliers d'années. Or, un document d'archive (Salmon, 1770) nous permet d'illustrer le type d'événements hydrologiques qui peuvent être responsables de telles remobilisations d'anciens stocks sédimentaires. D'après cette carte (Fig. 5.4), le 'déluge de Sainte Anne' aurait provoqué de forts ravinements sur le versant sud de la vallée de la Cleurie, ayant très certainement affecté le complexe morainique du Tholy. La plaine alluviale aurait alors été recouverte par des épaisseurs de sédiments allant jusqu'à 1,6 m. Or, le complexe morainique du Tholy a été déposé par le glacier de la Moselotte, en remontant la vallée de la Cleurie (Seret, 1967; Flageollet, 2002), ce qui amène à considérer que l'histoire d'exposition de son matériel puisse être très complexe. Nous pouvons alors supposer que le ratio $^{26}\text{Al}/^{10}\text{Be}$ mesuré dans les sédiments du lit de la Cleurie, le plus bas mesuré dans les Vosges (i.e. 1.4), puisse être causé par ce type d'événements, dont l'occurrence est tout à fait plausible dans d'autres bassins versants vosgiens.

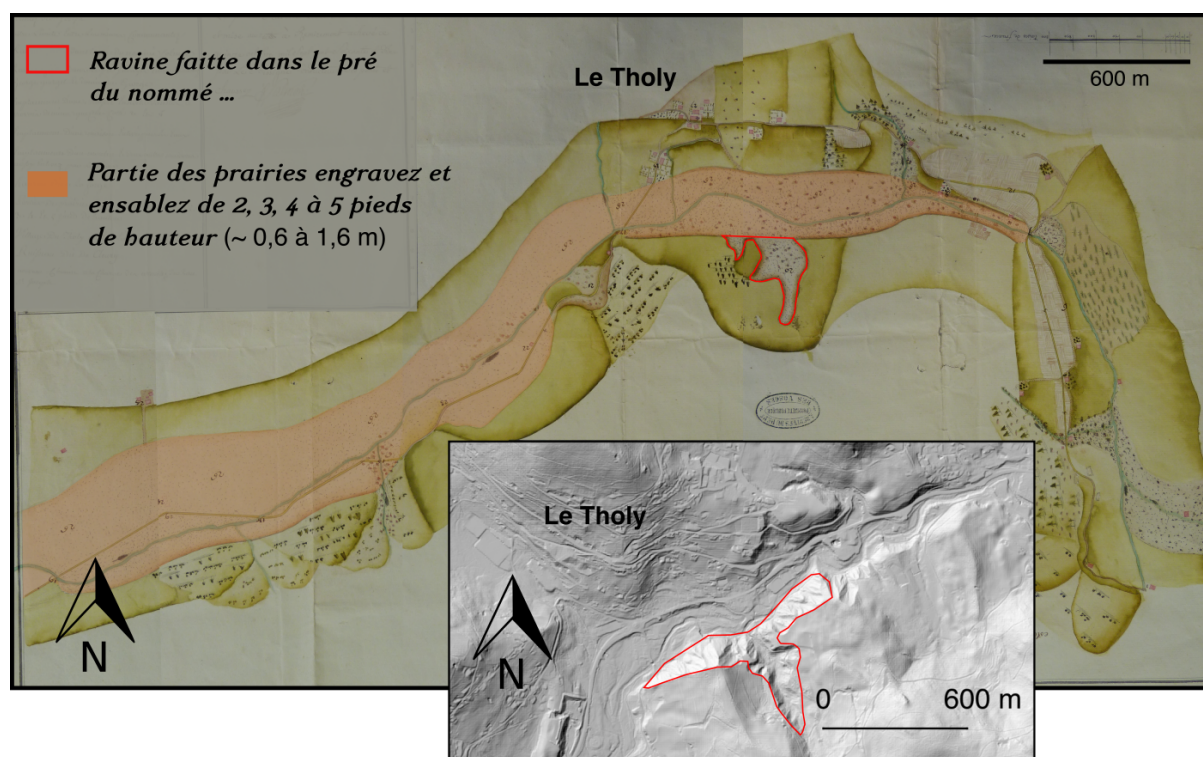


FIGURE 5.4: Extrait d'une carte dressée par le service des Ponts et Chaussées (Salmon, 1770), qui met en évidence les dégâts causés par 'le déluge de Sainte Anne' dans la vallée de la Cleurie, survenu en juillet 1770. L'extrait du LiDAR suppose que les 'ravines' ont eu lieu dans le complexe morainique du Tholy (Fig. 3.5B)

C'est précisément sur le bassin versant de la Cleurie qu'a porté l'étude du Chapitre 3. La partie suivante vise ainsi à synthétiser les résultats obtenus concernant l'étude de l'évolution fluviale post-DMG du bassin versant de la Cleurie, et à présenter quelques perspectives de recherche.

5.2 Reconstitution des ajustements fluviaux post-DMG du bassin versant de la Cleurie

À l'échelle du massif des Vosges, le bassin versant de la Cleurie présente le déséquilibre cosmogénique le plus marqué, avec un ratio $^{26}\text{Al}/^{10}\text{Be}$ très faible (1.4), indiquant un temps d'enfouissement apparent des sédiments de 3.4 Ma. Si les environnements glaciaires sont connus pour présenter des dynamiques de transfert de sédiments complexes (Porter et al., 2019), cette valeur mesurée à partir de sédiments sableux prélevés dans le lit actuel de la Cleurie témoigne d'une dynamique particulièrement complexe. A ce titre, l'étude du Chapitre 3 a visé à (i) évaluer l'état d'équilibre ou de déséquilibre actuel du système fluvial de la Cleurie et à (ii) reconstruire l'évolution fluviale post-DMG de la vallée. Nous présentons ici une synthèse des principaux résultats de notre étude.

L'englacement de la vallée de la Cleurie, marqué par les glaciers de la Vologne, de la Moselotte et du Cellet (Fig. 3.5), a conduit à la formation de différents niveaux glacio-lacustres (Flageolet et al., 1971), associés aux niveaux successifs des paléolacs :

- Niveau supérieur (~665 m) : Au maximum glaciaire, le lac est contenu à l'est par le glacier de la Vologne (complexe morainique du Bas-Beillard), à l'ouest par le glacier de la Moselotte (complexe morainique de Rain Brice) et au sud par le glacier du Cellet (Fig. 3.5). Suite à la première phase de retrait des glaciers, les dépôts du niveau supérieur sont formés depuis l'aval du complexe morainique du Tholy jusqu'à l'aval de l'actuel lac de Gérardmer (Fig. 3.11). Jusque là, le col de Rechaucourt servait d'exutoire au lac.
- Niveaux intermédiaire (~616 m) et Bas-Beillard (~598 m) : Ces dépôts correspondent à des phases de stabilité du lac lors de la fonte progressive des glaces. Les dépôts s'étendent du complexe morainique du Costet à celui du Tholy, puis du Bas-Beillard au Tholy (Fig. 3.11), avec un changement d'exutoire vers l'aval de la vallée de la Cleurie, entre les versants et le glacier de la Moselotte.
- Niveau inférieur (~533 m) et terrasses fluvio-glaciaires : Avec la poursuite du retrait du glacier de la Moselotte, un quatrième niveau de stabilité du lac (niveau inférieur) s'est formé en aval du complexe morainique du Tholy, probablement dans le même temps que les terrasses de la 'Basse Cleurie' formées par les eaux de fonte (Fig. 3.5 et 3.11).

Le retrait complet du glacier de la Moselotte a finalement permis à la Cleurie d'inciser le complexe morainique du Tholy, initiant le réajustement fluvial de la vallée. La partie inférieure de la vallée a commencé à se réajuster, menant à la formation de surfaces alluviales présentes tant dans la partie aval qu'amont de la vallée. La terrasse inférieure ($\sim 3,3$ m au-dessus de la plaine alluviale; Fig. 3.5) reflète ainsi une phase de stabilisation du fond de la vallée suivie d'un épisode d'incision, conduisant à la formation de la plaine alluviale moderne (Fig. 3.5).

L'étude du Chapitre 3 révèle également un déséquilibre fluvial très marqué, illustré par plusieurs aspects :

- **Étêtement du bassin versant** : Le retrait du glacier de la Vologne a redirigé l'exutoire du lac de Gérardmer vers la Vologne, réduisant la surface du bassin versant de la Cleurie d'environ 60 km^2 . Cela a considérablement diminué la puissance spécifique de la rivière, limitant sa capacité à mobiliser les matériaux glaciaires et à ajuster son profil en long.
- **Ruptures de pente** : Quatre ruptures de pente le long du profil du chenal principal mettent en évidence un déséquilibre marqué (Fig. 3.6). Le 'Saut de la Cuve' témoigne probablement d'une variation du niveau de base à la confluence avec la Moselotte, certainement liée à une phase d'incision de cette dernière. Des datations OSL d'une terrasse alluviale de la Moselotte (19 – 22 ka ; Cordier, 2015) situées 5 km en aval de la confluence avec la Cleurie confortent cette hypothèse. Une phase d'érosion régressive aurait donc été amorcée postérieurement au DMG. Cependant, la relative faible puissance spécifique de la Cleurie, amoindrie par l'étêtement de son bassin versant, n'a pas permis 'd'effacer' le knickpoint majeur du 'Saut de la Cuve'. Le gradient de déséquilibre des profils en longs des tributaires, croissant vers l'amont (depuis des profils ajustés ou droits vers des profils à knickpoints ou vallées suspendues), renforce l'hypothèse d'un signal d'ajustement fluvial amorcé après le DMG, mais dont la propagation vers l'amont a été considérablement entravée, une partie du signal ayant été potentiellement stoppée au niveau du 'Saut de la Cuve'.
- **Augmentation vers l'aval de la granulométrie de surface du fond du lit** : Une granulométrie fine en amont et grossière en aval de la vallée jusqu'à l'exutoire (Fig. 3.6C) indique un apport local massif de particules grossières dues à l'érosion latérale active du chenal, remaniant les dépôts du complexe morainique du Tholy.

Enfin, les séquences stratigraphiques de Morte Femme (MF ; Fig. 3.7) et Rain Brice (RB ; Fig. 3.8), localisées dans la zone anciennement submergée par les paléolacs, fournissent des premières contraintes chronologiques importantes. Les rythmites lacustres du profil MF datées de 10575-10281 à 9528-9294 cal BP suggèrent une sédimentation lacustre persistante jusqu'au début de l'Holocène. Dans l'unité alluviale sus-jacente, les datations obtenues (13295-13092 à

11870-11355 cal BP) sont en inversion stratigraphique par rapport aux rythmites lacustres, ce qui suggère une probable remobilisation du matériel daté (charbons). Cependant, ces dates permettent de donner un âge maximum d'une part pour la période de mise en place de la terrasse inférieure (Fig. 3.5) et d'autre part pour la fermeture du paléolac. Enfin, la datation obtenue dans le profil RB suggère une période de mise en place de la plaine inondable (Fig. 3.5) aux alentours de 1730-1543 cal BP.

A l'échelle de la partie du massif des Vosges anciennement englacée, en particulier le sud-ouest (Fig. 5.1), il conviendrait d'étendre l'analyse des profils en longs des réseaux hydrographiques dans d'autres bassins versants anciennement englacés. Cela permettrait de vérifier dans quelle mesure le déséquilibre fluvial observé dans la Cleurie est particulièrement prononcé en comparaison de ces derniers.



FIGURE 5.5: Photographie du front de taille du dépôt 'Le Passage', identifié comme une terrasse de kame. Auteur : Dominique Harmand (2008). Le rectangle rouge correspond à une échelle humaine. Le dépôt fait environ 20 m d'épaisseur.

A l'échelle du bassin versant, les perspectives de recherche viseraient à caractériser, et idéalement dater, les dépôts glacio-lacustres identifiés par Flageollet et al. (1971). Ce travail a été initié dans un de ces dépôts, qui correspond au dépôt dit 'Le Passage' (Fig. 5.5). Ce dernier a été identifié par Salome (1968) et Flageollet (1988) comme étant une terrasse de kame. Il fait partie du niveau inférieur de la classification des dépôts glacio-lacustres par altitude proposée par Flageollet et al. (1971) et est localisé juste en aval du complexe morainique du Tholy (Fig. 3.5B), entre le versant et le fond de vallée actuel. Il correspondrait donc à des dépôts formés durant la dernière phase de retrait du glacier de la Moselotte. Ainsi, sa datation permettrait de faire le lien entre le retrait des glaciers et le début du réajustement fluvial de la Cleurie. Un total de six logs ont été réalisés dans le front de taille du dépôt (Fig. 5.6),

dans l'unité supérieure du dépôt. Ils témoignent d'une très forte variabilité en termes de faciès d'écoulements. Les différentes unités correspondent à une alternance entre des dépôts sableux hétérométriques non structurés et des unités fluviales, voire à des paléochenaux.

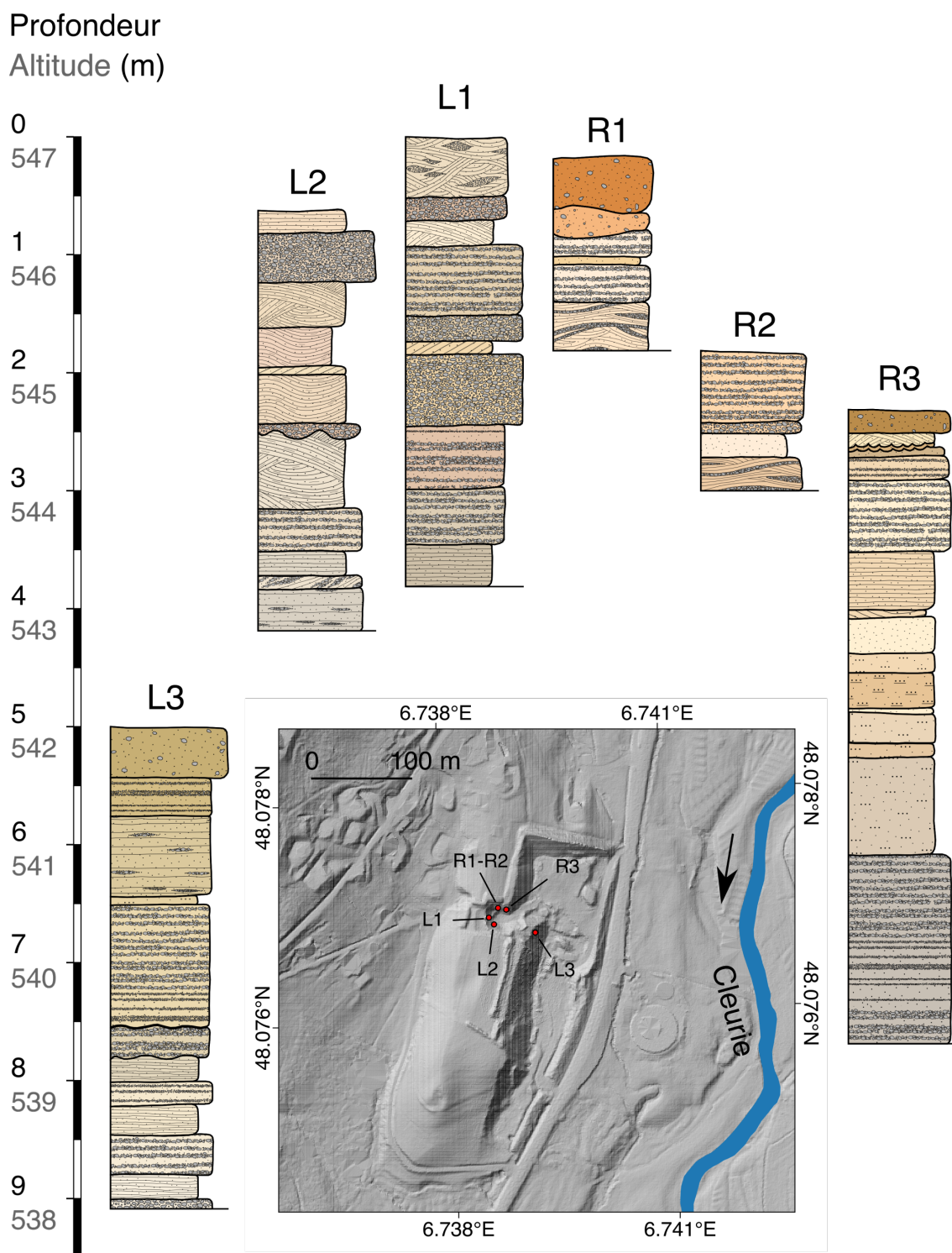


FIGURE 5.6: Localisation et logs des six profils stratigraphiques étudiés au sein du dépôt 'Le Passage' (PA). Les logs ont été réalisés par Coraline Fuchs (Fuchs, 2022).

5.3 Impact de l'anthropisation sur la trajectoire géomorphologique de la Fecht, de la Moselotte et de la Vologne

L'étude du Chapitre 4 a permis de dresser un portrait de la pression anthropique exercée au cours des deux derniers siècles sur les versants, fonds de vallées et lits fluviaux de la Fecht, de la Moselotte et de la Vologne. D'abord, bien que les trois hydrosystèmes fluviaux aient tous été considérablement aménagés au cours du XIX^{ème} siècle, un gradient de pression actuellement exercée par les structures hydrauliques (seuils, ponts, digues, etc.) a été mis en évidence. La Fecht est de loin la plus aménagée, avec environ deux fois plus de seuils (n=125) encore présents sur son cours que la Moselotte (n=59) et la Vologne (n=55; Fig. 4.9). Ensuite, une différence de temporalité concernant les périodes de constructions des seuils a été relevée entre les trois hydrosystèmes fluviaux. Le premier pic de pression humaine lié à ces aménagements au XIX^{ème} siècle semble avoir été légèrement plus précoce et plus court pour la Fecht que pour la Moselotte et la Vologne. Il a été suggéré que ce décalage puisse avoir été causé par l'histoire singulière du massif des Vosges, où la position de la frontière franco-allemande a fréquemment changé au cours des deux derniers siècles.

De manière générale, nous avons démontré que la pression humaine exercée depuis le XIX^{ème} siècle sur les systèmes fluviaux étudiés a joué un rôle important sur la trajectoire géomorphologique des lits fluviaux. Une impressionnante métamorphose fluviale de la partie aval de la Fecht a notamment été mise en évidence, directement liée à la construction de digues le long de son cours (Fig. 4.13A). Cependant, la comparaison de la trajectoire géomorphologique des trois hydrosystèmes fluviaux, en réponse à la pression humaine exercée au cours des deux derniers siècles à l'échelle des versants, fonds de vallées et lits fluviaux, a également permis de souligner l'importance cruciale de la prise en compte du contexte géomorphologique pour comprendre la dynamique fluviale passée et actuelle. En effet, la morphodynamique de ces trois cours d'eau, bien que significativement impactée par l'anthropisation, est également fortement contingentée par les facteurs de contrôle morpho-structuraux, comme la largeur et la pente du fond de vallée, ou encore les héritages (fluvio-)glaciaires présents dans le fond de vallée.

Enfin, la démarche géohistorique que nous avons appliquée dans le cadre de l'étude du Chapitre 4 a bien illustré la nécessité d'appréhender les hydrosystèmes étudiés, dont le fonctionnement est à l'interface entre un système naturel et un système social, comme des 'anthroposystèmes' (e.g. Valette and Carozza, 2010; Schmitt, 2011; Lespez and Dufour, 2021). A ce titre, dans une optique de meilleure gestion voire de restauration fonctionnelle des hydrosystèmes, il convient ni de définir, ni de viser un quelconque état de 'référence' en lien avec le fonctionnement passé de ces derniers. Nous avons vu en effet que la dynamique fluviale

observée au XIX^{ème} siècle était largement conditionnée par un contexte particulier (combinaison du minimum forestier et du pic du PAG), qui fait lui-même suite à une trajectoire passée pluri-séculaire (voire pluri-millénaire), sous contrôle anthropique et climatique (Figs. 4.6 et 4.12). Or, le contexte est aujourd'hui bien différent et est amené à continuer de changer rapidement dans un futur proche.

Les perspectives développées dans le Chapitre 4 visent à mieux déconvoluer les facteurs de contrôle anthropiques des facteurs de contrôle morpho-structuraux, climatiques et hydrologiques. Dans le cadre de l'élaboration de stratégies de gestion, voire de restauration fonctionnelle des hydrosystèmes étudiés, à travers notamment le recouvrement de leur mobilité latérale et de la connectivité hydro-sédimentaire longitudinale, il conviendrait désormais d'étudier plus en détail le potentiel de remobilisation des stocks sédimentaires des fonds de vallées. Cela implique d'étudier (i) la localisation et la composition des stocks sédimentaires, (ii) la dynamique longitudinale du transport de la charge de fond et (iii) l'évolution longitudinale de la granulométrie de la charge de fond.

La section suivante vise à présenter les résultats préliminaires obtenus dans le cadre de cette thèse concernant les aspects (i) et (ii) et à illustrer l'intérêt de leur usage. Les aspects (iii) ont quant à eux déjà été abordés dans le Chapitre 4.

5.4 Vers un diagnostic géomorphologique fonctionnel pour une meilleure gestion des rivières vosgiennes

5.4.1 Apports de l'étude des stocks sédimentaires de fonds de vallée : l'exemple de la Fecht

Le remplissage sédimentaire du fond de vallée de la Fecht et son évolution longitudinale ont été étudiés à travers quatre profils stratigraphiques, distribués d'amont en aval depuis la tête de bassin anciennement englacé jusqu'au cône alluvial développé par le cours d'eau à son débouché dans le Fossé Rhéna (Fig. 5.7). Ils correspondent à deux affleurements de terrasses (LUT et WV), actuellement déconnectés du chenal actif, et à deux profils de berges (MET et AME). Ces résultats ont été acquis dans le cadre du stage de Master 2 de Margaux Claudepierre (Claudepierre, 2023). Les quatre profils stratigraphiques sont présentés ci-après, d'amont en aval.

Le profil MET est localisé en tête de bassin versant, dans la zone anciennement englacée (Fig. 5.7). Il correspond à une encoche d'érosion dans la berge active de la Fecht et représente probablement l'un des derniers sites de berge non-aménagée de la Fecht dans sa partie amont (Fig. 5.8). Il est donc très informatif, dans la mesure où il illustre le matériel potentiellement

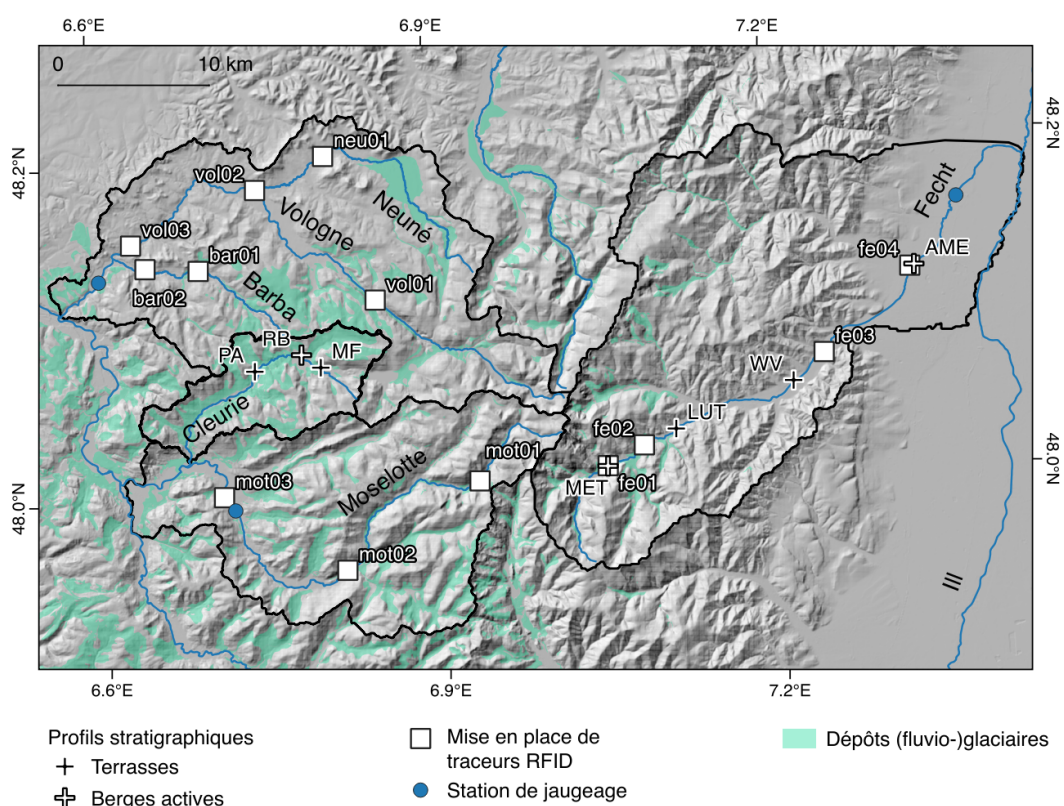


FIGURE 5.7: Localisation des (1) sites de mise en place des traceurs RFID dans les lits de la Fecht, Moselotte, Vologne, Neuné et Barba et des (2) profils stratigraphiques étudiés dans les fonds de vallée de la Fecht et de la Cleurie.

remobilisable par le cours d'eau dans son secteur amont. D'une hauteur de 1.20 m, il est constitué d'alluvions très grossières et bien émoussées avec des diamètres apparents les plus longs (Longest Visible Axis; LVA; e.g. Garefalakis et al., 2023) de particules allant de 100 à plus de 1000 mm dans sa partie basale (Fig. 5.8D). La présence d'encoches d'érosion (Fig. 5.8B) témoigne effectivement d'une érosion latérale de la berge, dont nous supposons que le taux de recul est relativement faible, au vu de la granulométrie très grossière. Cependant, il conviendrait de déterminer dans quelle mesure la compétence actuelle du cours d'eau permet ou non le transport de telles particules, dès lors qu'elles se détachent de la berge et rejoignent le fond du lit. Le suivi des traceurs RFID (site fe01; Fig. 5.7) devrait permettre à terme de vérifier cela. En effet, la granulométrie des particules équipées de traceurs RFID a été sélectionnée de telle manière à être représentative de la charge de fond du lit. En l'occurrence, pour le site fe01 situé directement au pied du profil MET (Fig. 5.7), le D_{50} et le D_{90} de l'axe-b des 100 particules mesurées (Fig. 4.12C) ont une valeur de 137 et 261 mm, respectivement. Nous y avons également mesuré plusieurs particules dont l'axe-b est supérieur à 500 mm (dont un bloc de 830 mm). Si la granulométrie de la charge de fond du lit semble bien correspondre à la granulométrie apparente de la berge, nous ne sommes pas en mesure pour le moment de déterminer si la charge de fond provient ou non de la berge.

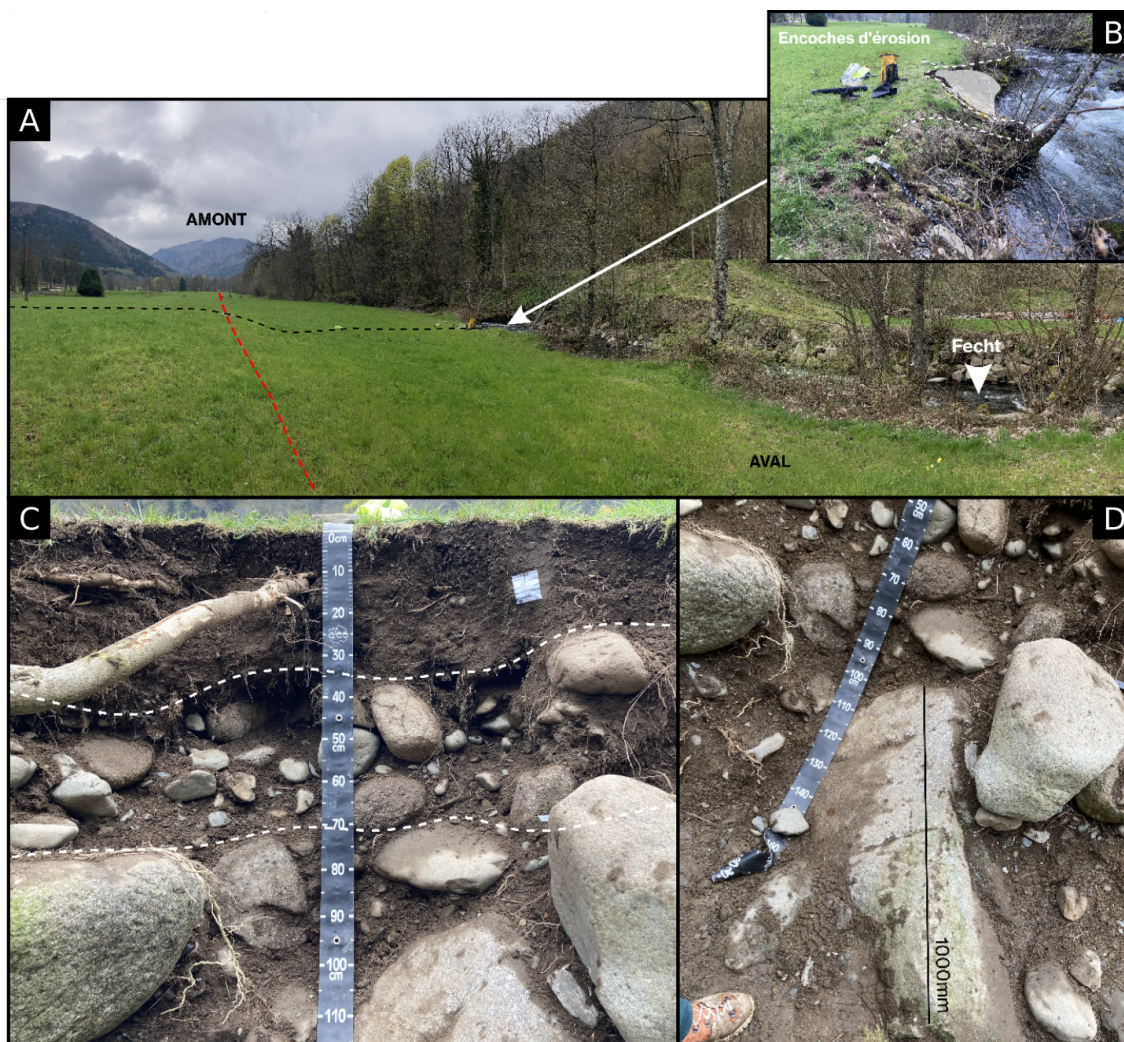


FIGURE 5.8: Contexte géomorphologique et photographie du profil de berge MET, correspondant à une encoche d'érosion dans une berge de la Fecht.

Le profil LUT est localisé environ 5 km en aval du profil MET, en aval immédiat de la zone anciennement englacée (Fig. 5.7). Il correspond à un talus de terrasse, d'une hauteur d'environ 6 m, situé 20 m au nord du lit actif de la Fecht (Fig. 5.9). Au sein d'une matrice sablo-limoneuse, les éléments grossiers du profil LUT1 ont des LVA compris entre 50 et 170 mm dans sa partie supérieure et entre 200 et 500 mm dans sa partie inférieure. Les datations par radiocarbone effectuées sur des charbons prélevés dans les unités graveleuses du profil LUT2 (Fig. 5.9C), indiquent que le lit actif de la Fecht venait saper la base de la terrasse, jusqu'à la fin du XX^{ème} siècle. Les éléments grossiers du profil LUT2 ont des LVA compris entre 100 et 200 mm. Ils présentent donc une granulométrie inférieure à celle de la base du profil LUT1. Concernant la granulométrie de la charge de fond du lit actif, situé à 20 m du profil LUT, le D_{50} et le D_{90} de l'axe-b des 100 particules mesurées (Fig. 4.12C) ont une valeur de 130 et 261 mm, respectivement. La granulométrie de la charge de fond du lit serait donc légèrement plus grossière que celle mesurée dans le profil LUT2, à la base de la terrasse.

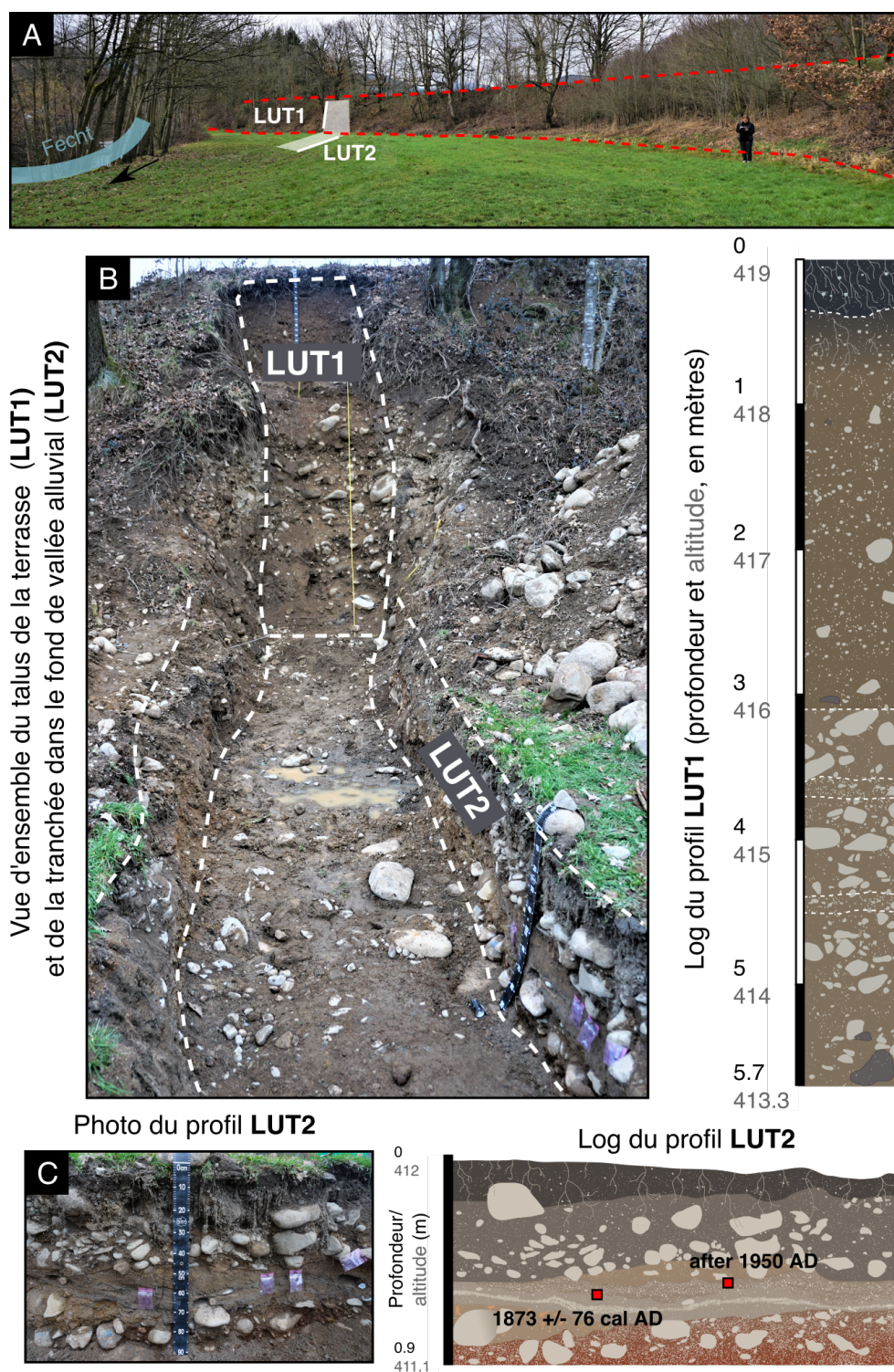


FIGURE 5.9: Profil stratigraphique LUT, correspondant à une terrasse et au fond de vallée alluvial de la Fecht. (A) Position du profil étudié par rapport au chenal actuel. (B) Photographie et log du profil LUT1, correspondant au talus de la terrasse. (C) Photographie et log du profil LUT2, correspondant au prolongement dans le fond de vallée alluvial. Les logs ont été réalisés par Claudepierre (2023).

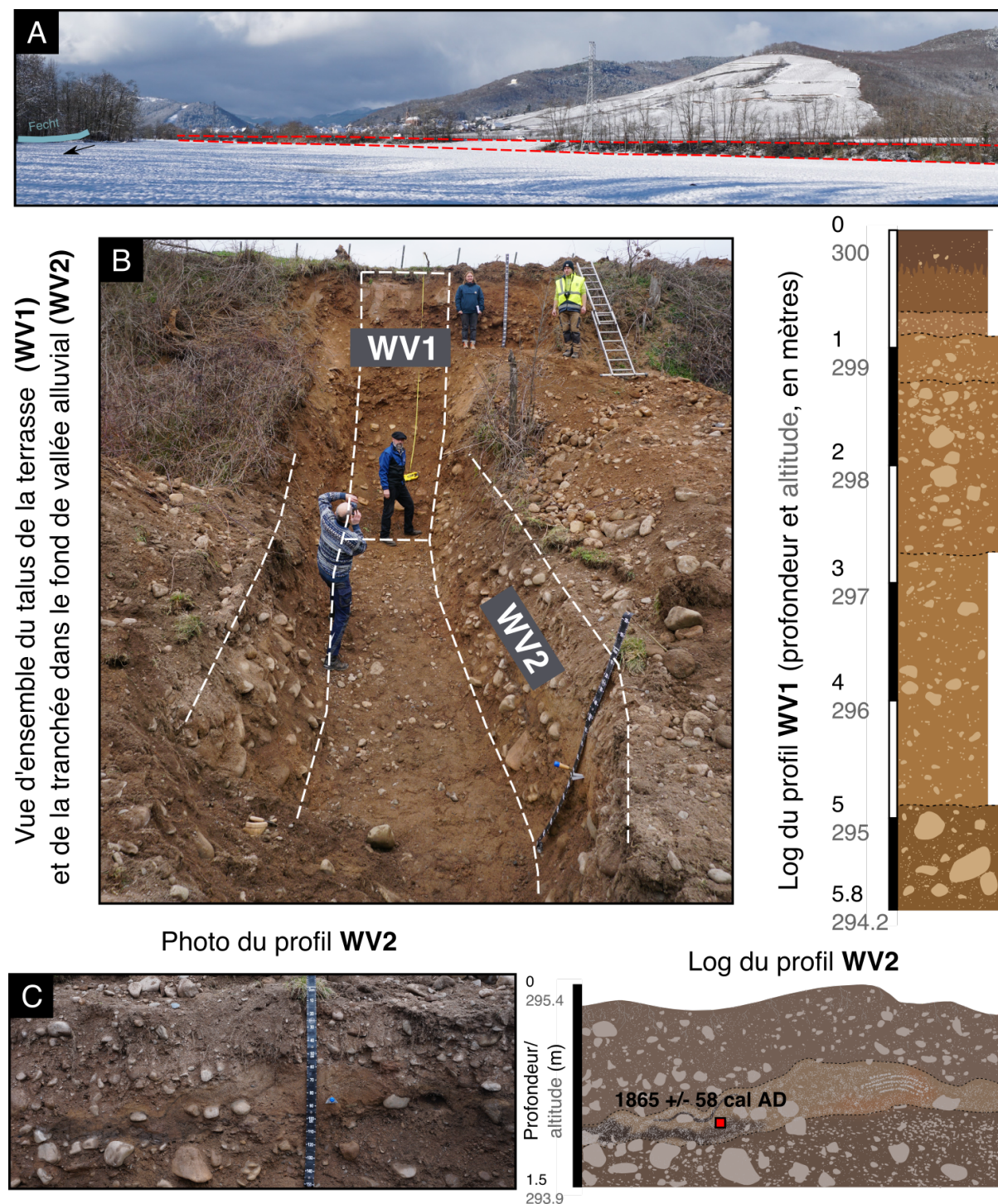


FIGURE 5.10: Profil stratigraphique WV, correspondant à une terrasse et au fond de vallée alluvial de la Fecht. (A) Position de la terrasse étudiée par rapport au chenal actuel. (B) Photographie et log du profil WV1, correspondant au talus de la terrasse. (C) Photographie et log du profil WV2, correspondant au prolongement dans le fond de vallée alluvial. Les logs ont été réalisés par Claudepierre (2023).

Le profil WV est localisé environ 9 km en aval du profil LUT, dans la zone où la vallée est considérablement plus large (~ 1000 m; Fig. 5.7). Il correspond à un talus de terrasse, d'une hauteur d'environ 6 m, situé 130 m au Nord du lit actif de la Fecht (Fig. 5.10). Au sein d'une matrice limono-sableuse, les éléments grossiers du profil WV1 ont des LVA compris entre 100 et 200 mm dans sa partie inférieure. La datation par radiocarbone effectuée sur un charbon prélevé dans le profil WV2 (Fig. 5.9C), indique que le lit actif de la Fecht venait saper la base de la terrasse, jusqu'à la fin du XIX^{ème} siècle, ce qui est cohérent avec la position du chenal relevée depuis la carte d'état-major (cf. Chapitre 4). Les éléments grossiers du profil WV2 ont des LVA compris entre 100 et 200 mm, ce qui est comparable à la granulométrie de WV1. Concernant la granulométrie de la charge de fond du lit actif, situé à 130 m du profil WV, le D_{50} et le D_{90} de l'axe-b des 100 particules mesurées (Fig. 4.12C) ont une valeur de 100 et 201 mm, respectivement.

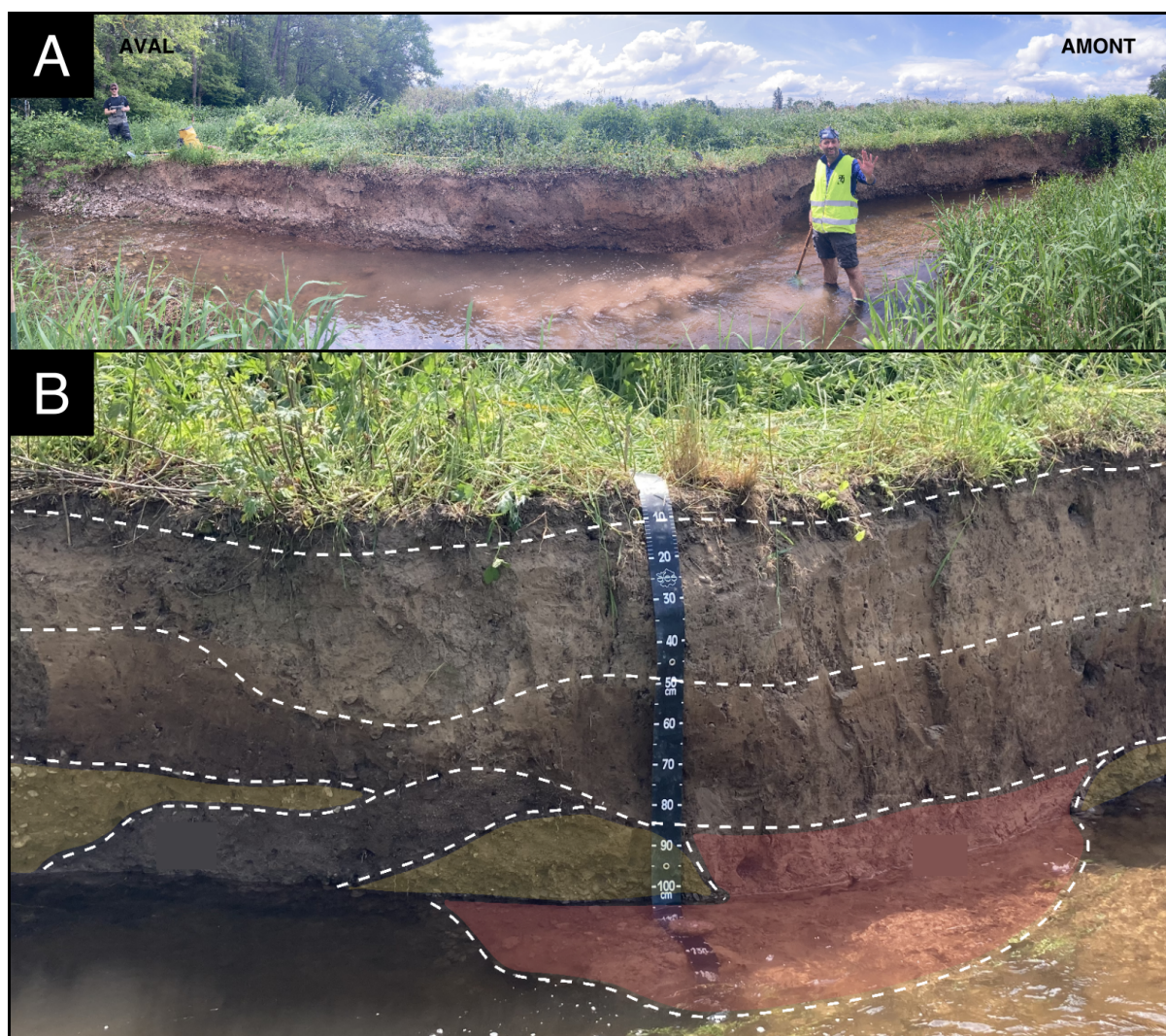


FIGURE 5.11: Profil stratigraphique AME, correspondant à une berge concave de la Fecht. (A) Vue d'ensemble de la berge concave. (B) Vue rapprochée du profil de berge. L'ensemble du profil est limono-sableux à grano-classement positif, à l'exception des unités jaune et rouge. Elles correspondent respectivement à des unités gravo-sableuses et argileuses.

Le profil AME est localisé environ 12 km en aval du profil WV, sur le cône alluvial au débouché du massif Fig. 5.7), dans le secteur qui présentait un style en anabranches au milieu du XIX^{ème} siècle (Fig. 4.13A). Il correspond à une berge concave de la Fecht, d'une hauteur d'environ 1.20 m, et représente également un des rares tronçons non aménagés de la Fecht (Fig. 5.11). Le profil est de composition essentiellement limono-sableuse, à grano-classement positif. A sa base, une unité gravillonneuse (aplat jaune sur la Figure 5.11) vient recouper une unité limono-argileuse (aplat rouge sur la Figure 5.11). Concernant la granulométrie de la charge de fond du lit actif, le D₅₀ et le D₉₀ de l'axe-b des 100 particules mesurées (Fig. 4.12C) ont une valeur de 65 et 105 mm, respectivement.

Ce type de profils stratigraphiques fournit des informations précieuses concernant les stocks sédimentaires présents dans la vallée, et l'évolution de leur composition d'amont en aval. En l'état actuel de l'étude, il est difficile de tirer des conclusions définitives quant au potentiel de remobilisation de ces stocks sédimentaires. Cependant, ces premières analyses révèlent d'ores et déjà plusieurs informations importantes. Premièrement, le remplissage du fond de vallée en tête de bassin (profil MET) est très grossier, avec des LVA pluri-décimétriques à métriques. Il est donc possible d'en déduire que la mobilité latérale de la Fecht dans ce secteur est morphologiquement très contrainte. Deuxièmement, dans la partie intermédiaire de la vallée, les datations par radiocarbone effectuées dans les profils LUT2 et WV2 révèlent une mobilité latérale de la Fecht active jusqu'au XIX^{ème} et XX^{ème} siècle. Troisièmement, la composition des unités dans lesquelles ont été prélevés les échantillons radiocarbone (LVA des galets compris entre 100 et 200 mm), nous donne un premier élément concernant la compétence de la Fecht sur cette même période. Or, nous avons pu mesurer que la charge de fond du lit de la Fecht (en contrebas des profils LUT et WV) contient 50% de particules dont l'axe-b est supérieure à 130 et 100 mm, respectivement. Cependant, la compétence actuelle de la Fecht devra être mesurée pour déterminer le potentiel de mise en mouvement de telles particules. En effet, nous avons montré que les conditions hydro-climatiques et d'occupation du sol au XIX^{ème} siècle étaient très différentes du contexte actuel (cf. Chapitre 4). Nous faisons d'ailleurs l'hypothèse que la compétence actuelle de la Fecht est moins importante que celle du XIX^{ème} siècle.

Ces données illustrent donc l'importance de la prise en compte de la composition des stocks sédimentaires, qui constituent les héritages (fluvio-)glaciaires des fonds de vallées, et qui conditionnent actuellement la dynamique fluviale des rivières vosgiennes. D'une part, ce type d'analyses stratigraphiques des stocks sédimentaires des fonds de vallées devrait être approfondi au niveau de la Fecht, en tentant notamment de dater les dépôts de terrasses (*via* par exemple des profils de concentration en nucléides cosmogéniques ou luminescence) dans la mesure où ils constituent des unités morpho-stratigraphiques/héritages fluviaux de premier plan. D'autre part, ce type d'études stratigraphiques devra être transposée sur le versant ouest du massif des Vosges, dans les vallées de la Moselotte et de la Vologne.

5.4.2 Transport de la charge de fond dans les trois hydrosystèmes étudiés

Afin de caractériser la dynamique longitudinale du transport de la charge de fond des lits fluviaux de la Fecht, de la Moselotte et de la Vologne, un total de 1350 traceurs RFID (e.g. Gilet et al., 2020; Peeters, 2021; Liébault et al., 2024) ont été mis en place sur 13 sites (Fig. 5.7), entre mars et octobre 2021 (Fig. 5.12). Une partie des résultats présentés dans cette section ont été acquis dans le cadre des stages de Master de Alexis Cervantes (Cervantes, 2021) et Romain Soeur (Soeur, 2021). Trois des sites sont situés sur des affluents de la Vologne : le Barba et le Neuné. Ils ont été équipés dans le cadre d'une collaboration avec la Société d'Histoire Naturelle et d'Ethnographie de Colmar (SHNEC), qui pilote un projet de réintroduction de la Mulette perlière (*Margaritifera margaritifera*) dans le bassin de la Vologne (Umbrecht, 2021). Des précisions et des résultats préliminaires concernant cette collaboration sont apportés en Annexe C.2.

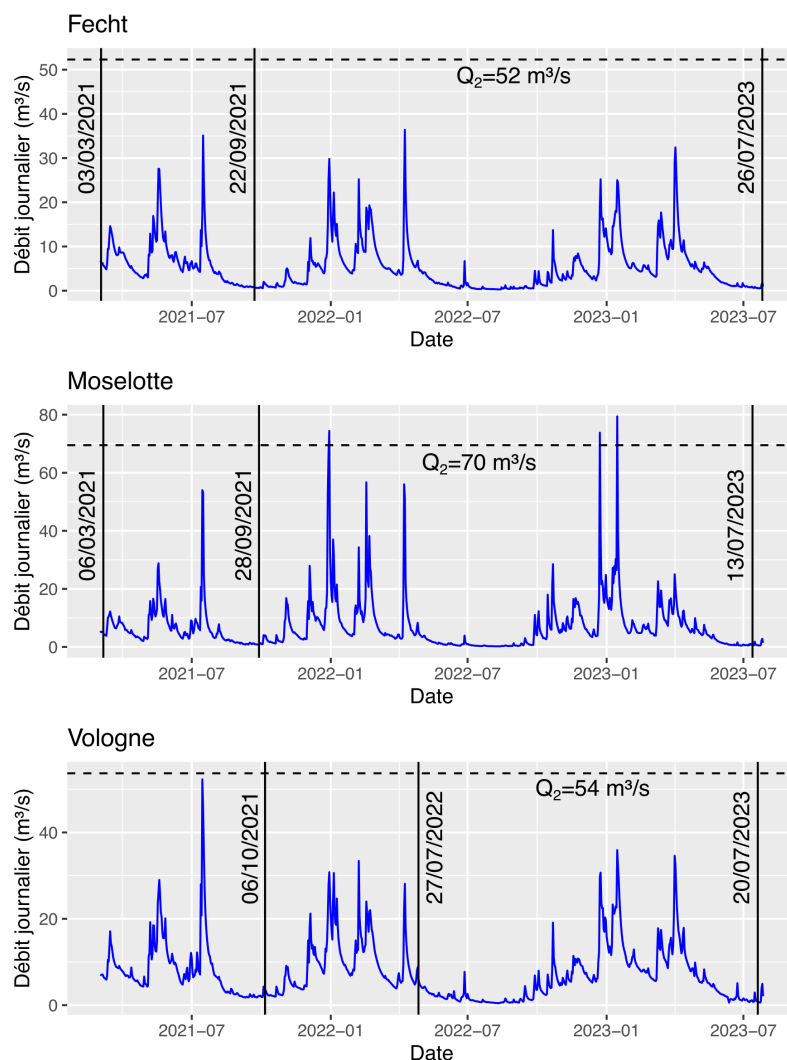
Deux premiers levés ont été réalisés sur l'ensemble des 13 sites (Fig. 5.12). Les périodes couvertes par les deux levés furent relativement sèches. Seule la Moselotte a vu son débit dépasser Q_2 , sur la deuxième période. Malgré une deuxième période de suivi plus longue et dont les débits journaliers maximums atteints ont été légèrement supérieurs à ceux atteints pendant la première période (Fig. 5.12), la proportion de traceurs mobiles durant la deuxième période est nettement inférieure à celle mesurée durant la première période. Cela témoigne d'un biais bien connu lié à la surexposition des particules lors de leur mise en place initiale, qui engendre une mobilité artificielle (e.g. Houbrechts et al., 2012; Liébault et al., 2024). Dans notre cas, l'impact de la surexposition des particules sur la mobilité artificielle engendrée durant la première période de suivi semble particulièrement significatif, ce qui implique nécessairement un troisième levé avant toute comparaison et interprétation des résultats obtenus. De plus, depuis le dernier levé, les pics de crues ont été bien plus élevés sur les trois cours d'eau. La Fecht a dépassé une fois Q_5 et deux fois Q_2 . La Moselotte a dépassé une fois Q_{10} et une fois Q_5 . La Vologne a dépassé trois fois Q_2 .

Un levé supplémentaire (minimum) apportera donc de précieuses informations quant à la dynamique longitudinale du transport de la charge de fond sur les trois cours d'eau étudiés, en particulier dans notre cas lors de crues clairement morphogènes. En croisant les résultats avec la granulométrie du fond du lit, dont la mesure a déjà été effectuée (Fig. 4.12C), les résultats devraient permettre d'évaluer l'impact des héritages glaciaires et fluviaux sur le transport de la charge de fond. Nous émettons notamment l'hypothèse que celui-ci est de type sélectif sur les linéaires de fonds de vallées glaciaires, quelles que soient l'intensité des crues. Les résultats préliminaires portant sur le deuxième levé semblent déjà indiquer un seuil de diamètre de particules (~ 150 mm), commun aux trois cours d'eau, au delà duquel les particules ne sont pas (Fecht et Moselotte) ou guère (Vologne) mobilisées (Fig. 5.13). Cependant, afin de pouvoir vérifier l'hypothèse énoncée plus haut, il conviendra d'étudier

la mobilité des particules au sein de chaque site étudié, pour comparer la dynamique du transport de la charge de fond entre les sites glaciaires et les sites alluviaux. Le nouveau levé, qui intégrera des crues bien plus importantes, sera d'autant plus instructif.

Hydrogrammes et périodes de suivi

Proportion de traceurs mobiles



Levé 1 : 92% (352/381)

Levé 2 : 24% (91/374)

Levé 1 : 93% (244/263)

Levé 2 : 27% (66/242)

Levé 1 : 49% (109/222)

Levé 2 : 26% (67/258)

FIGURE 5.12: Hydrogrammes des débits journaliers de la Fecht, Moselotte et Vologne depuis la mise en place des traceurs RFID jusqu'au 2^{ème} levé, aux stations de jaugeage localisées sur la Figure 5.7. Les valeurs de Q_2 ont été calculées sur les périodes 1972-2024, 1993-2024 et 1988-2024, pour la Fecht, Moselotte et Vologne, respectivement. La proportion de traceurs 'mobiles' à chaque suivi a été défini sur la base d'une distance de transport minimale de 5 m.

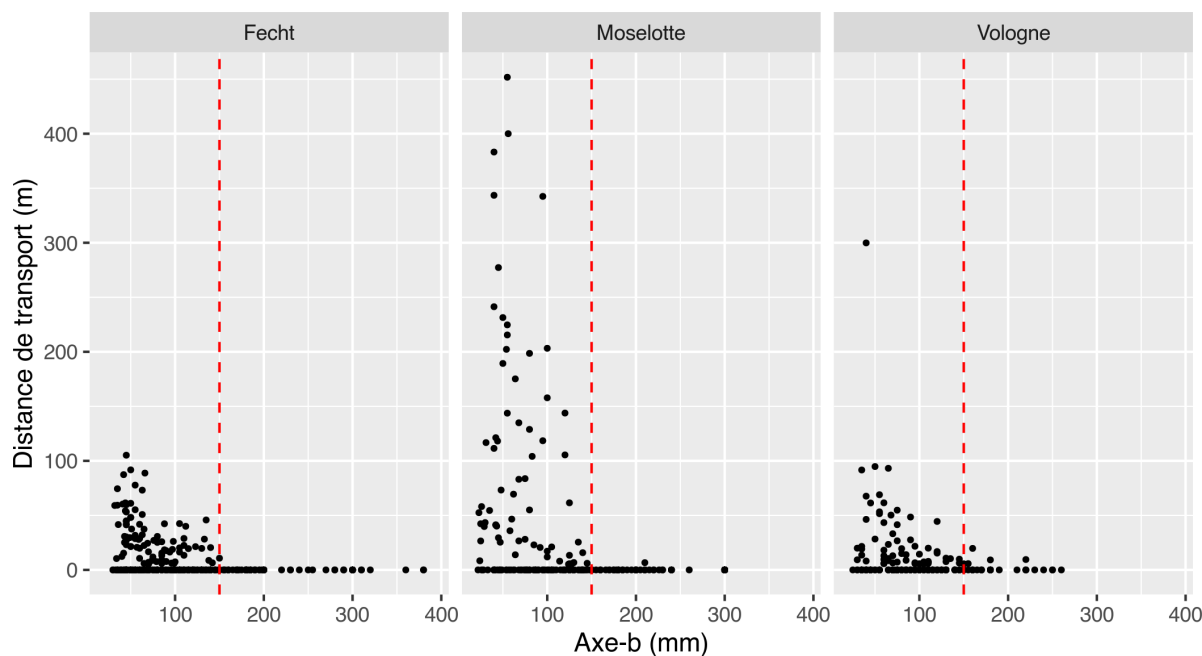


FIGURE 5.13: Distance de transport des traceurs RFID en fonction de leur axe-b, entre le 1^{er} et le 2^{ème} levé, pour chaque cours d'eau, tous sites confondus. Le trait pointillé rouge met en évidence un seuil de diamètre de particules (~ 150 mm), commun aux trois cours d'eau, au delà duquel les particules n'ont pas été mobilisées.

5.4.3 Perspectives en terme de gestion des rivières vosgiennes

Cette section finale vise à mettre en perspective des recommandations préliminaires de gestion, en se basant sur les résultats présentés dans ce travail de thèse.

Premièrement, il convient de souligner la forte pression anthropique qui affecte le fonctionnement des systèmes fluviaux vosgiens, qui doivent être considérés comme des 'anthroposystèmes' (cf. Chapitre 4; Lespez and Dufour, 2021). Deuxièmement, une importante spécificité des systèmes fluviaux vosgiens est leur hétérogénéité en terme de contexte géomorphologique, que ce soit à l'échelle de l'entière du massif (englacement partiel, cf. Chapitre 2) ou des hydrosystèmes étudiés dans ce travail (cf. Chapitre 4): Fecht, Moselotte et Vologne. En effet, nous avons montré que les lits de ces derniers évoluent à la fois dans des vallées glaciaires (amont) et alluviales (aval). Or, la gestion des cours d'eau doit tenir compte des conditions sédimentaires et géomorphologiques locales (Wohl et al., 2024), qui peuvent résulter d'héritages glaciaires (Mason and Polvi, 2023) et/ou de pressions anthropiques variées (Brown et al., 2018), desquels résultent la dynamique fluviale contemporaine. Sur la base des résultats obtenus dans cette étude, nous proposons dans cette section des recommandations opérationnelles visant à recouvrer autant que possible, et aussi durablement que possible, des processus morpho-sédimentaires libres (restauration fonctionnelle). Ceci doit notamment améliorer leur dynamique écologique tout en renforçant leur résilience face au changement climatique.

5.4.3.1 Prise en compte et caractérisation des héritages glaciaires

Concernant la prise en compte et la caractérisation des héritages (fluvio-)glaciaires, les recommandations que nous proposons, qui s'appliquent en particulier aux hydrosystèmes fluviaux drainant la partie méridionale du massif ('Hautes Vosges cristallines'), sont les suivantes :

- Tenir compte des héritages glaciaires dans la compréhension de la dynamique morpho-sédimentaire contemporaine des rivières;
- Cartographier précisément les héritages glaciaires (ex. moraines) et fluvio-glaciaires (ex. terrasse de kame), car ils constituent des stocks sédimentaires potentiels (voire partiels) pour les rivières ; les dépôts glaciaires non (totalement) remobilisables constituent quant à eux un facteur de diversification des habitats (voir aussi infra). Le croisement de la carte géologique harmonisée au 1/50 000 du BRGM, avec un MNT à résolution métrique (RGEAlti, IGN), constitue pour cela une bonne première approche. Une vérification de terrain postérieure est hautement souhaitable. Les bassins versants à faible ratio $^{26}\text{Al}/^{10}\text{Be}$ (cf. Chapitre 2) sont susceptibles de présenter des stocks sédimentaires conséquents, à prendre en compte dans les diagnostics morpho-sédimentaires;
- Caractériser les stocks sédimentaires et évaluer le volume de sédiments potentiellement remobilisables par les rivières vosgiennes. Cela requiert une cartographie des terrasses, associé à une analyse stratigraphique. La comparaison sédimentologique des terrasses avec la granulométrie de surface du fond des lits, couplée à l'analyse des résultats des traceurs RFID, doit permettre d'identifier les dépôts remobilisables. Cette approche pourra être consolidée avec l'intégration de l'évolution planimétrique historique, les enveloppes de mobilité maximales ne devant pas être considérées comme des cibles, mais pouvant permettre de confirmer les potentiels de remobilisation;
- Considérer les héritages (fluvio-)glaciaires comme un facteur de diversification des habitats aquatiques et riverains, comme le montre l'exemple de l'étude de la Mulette perlière (*Margaritifera margaritifera*), qui doit être considérée comme une espèce 'parapluie'. En effet, les héritages (fluvio-)glaciaires conditionnent également la qualité des habitats (et donc la présence) d'autres espèces, telles que la Truite (*Salmo trutta*), le Chabot (*Cottus gobio*), ainsi que des espèces d'écrevisses autochtones (*Austropotamobius pallipes*, *Astacus astacus*) et divers invertébrés (Umbrecht, 2021).

5.4.3.2 A l'échelle du bassin versant

L'ensemble de nos résultats montre l'importance de tenir compte de l'évolution pluri-séculaire de l'occupation du sol des bassins versants vosgiens comme facteur ayant fait fluctuer la

fourniture sédimentaire aux rivières et, par conséquent, l'intensité de leur mobilité latérale, voire leur style fluvial. Plus précisément, il est clair que les conditions hydro-climatiques (Petit Âge Glaciaire) et d'occupation du sol (minimum forestier) du milieu du XIX^{ème} siècle diffèrent significativement des conditions actuelles. Par conséquent, même en 'désaménageant' les rivières vosgiennes (Lespez et al., 2017) dans les secteurs sans enjeux humains, le style fluvial du XIX^{ème} siècle à tresses et/ou anabranches ne serait probablement pas recouvert, car les facteurs de contrôle de cette dynamique passée exacerbée n'existent plus aujourd'hui. Cet important résultat de la présente thèse devrait amener les gestionnaires à considérer des programmes de restauration ambitieux, car leur mise en œuvre devrait conduire à retenir des enveloppes de mobilité latérale des rivières bien plus étroites que dans le passé, ce qui atténue les risques d'érosion latérale non-maîtrisée. La question de la détermination de l'emprise des enveloppes de mobilité latérale en correspondance avec les facteurs de contrôle actuels (voire futurs) des bassins versants est une question ouverte qui dépasse le cadre de la présente recherche. Il nous semble toutefois que cela devrait passer par des actions de restauration itératives et adaptatives, accompagnées de suivis de terrain fins (Charles and Kalaora, 2003; Schmitt et al., 2013).

Plus globalement, nos résultats montrent très clairement (et dans notre cas pour la première fois dans les Vosges), à l'instar d'autres auteurs (e.g. Dufour and Piégay, 2009; Brown et al., 2018), que la notion d'état de référence de la morphologie et du fonctionnement morpho-sédimentaire des hydrosystèmes fluviaux n'a pas de validité scientifique. Tout programme de restauration morpho-sédimentaire doit donc essentiellement viser à recouvrer des processus fluviaux (inondation par surverse des berges, transport solide grossier, érosion des berges, dynamique latérale de la bande active avec rajeunissement des successions végétales, dépôt d'alluvions en lit mineur et majeur, etc.), plutôt que les formes du passé (Bravard, 2003; Schmitt, 2011; Schmitt et al., 2013; Lespez et al., 2015; Wohl, 2020; Gariépy-Girouard et al., 2024).

5.4.3.3 A l'échelle des fonds de vallées et des lits mineurs

A l'échelle des fonds de vallées et des lits mineurs, les recommandations de gestion que nous proposons sont les suivantes :

- Réévaluer, *a minima* localement, l'utilité de maintenir des digues construites depuis le milieu du XIX^{ème} siècle pour contracter l'espace de mobilité latérale. Sur cette base, il semble pertinent, localement, d'envisager l'arasement ou le déplacement de digues pour élargir l'espace de mobilité des rivières. Le recouvrement de la mobilité latérale est bénéfique pour la société. En effet, cela permet une meilleure adaptation au changement climatique (Buffin-Bélanger et al., 2015) par :

- un meilleur écrêtement des crues (Dister et al., 1990);
 - une meilleure recharge de la nappe phréatique, qui plus est en eau de bonne qualité (Carbiener and Trémolières, 2003), ce qui soutient les étiages et permet de lutter contre les assèchs, tout en favorisant le développement de la ripisylve, laquelle rafraîchit le climat local en été par amplification de l'évapotranspiration (Dunea et al., 2021). Le relèvement du niveau piézométrique peut également favoriser l'agriculture;
 - l'extension des refuges thermiques estivaux par les exfiltrations phréatiques, ce qui bénéficie aux écosystèmes aquatiques et riverains (Schmitt, 2011);
 - l'augmentation de la biodiversité et l'amélioration du fonctionnement des écosystèmes aquatiques et riverains, ce qui constitue également un facteur d'adaptation au changement climatique (Biron et al., 2018).
- Réévaluer la nécessité de certains seuils, notamment dans les secteurs glaciaires où leur maintien semble particulièrement discutable (selon toutefois les situations locales), mais cela concerne également certains linéaires alluviaux. Il peut ainsi être envisagé d'arasier certains seuils pour rétablir la continuité hydro-sédimentaire et écologique longitudinale (Melun, 2012), sans pour autant remettre en question la stabilité du profil en long en milieu alluvial (les linéaires glaciaires comprenant des blocs semi-métriques à métriques ne présentent *a priori* aucun risque d'instabilité du profil en long);
 - Réévaluer la pertinence de certaines protections de berges, qui sont très présentes sur les trois rivières étudiées, en particulier la Fecht, en tenant compte des héritages glaciaires, en vue de leur suppression locale quand cela n'impacte pas des enjeux locaux. La suppression (là aussi locale) de protections des berges sur les linéaires glaciaires (voire fluvio-glaciaires, selon les résultats des analyses sédimentologiques évoquées plus haut) ne devrait pas entraîner de risque notable d'érosion non-maîtrisée des berges, mais devrait au contraire constituer un facteur de diversification des habitats des berges et du fond des lits par la présence de blocs dépassant la compétence des rivières (Bunte and Abt, 2001), au sein d'une matrice hétérométrique plus fine.

5.4.3.4 Stratégie générale et approche adaptative

De façon globale, il nous semble essentiel d'adopter une stratégie adaptative, en mettant en œuvre des sites-tests (sans enjeux humains), à reproduire en fonction des retours d'expérience (Schmitt et al., 2013, 2019; Chardon et al., 2021). Sur un cours d'eau très fortement stabilisé comme la Fecht, l'exemple de la partie aval de cette rivière (tronçon s'écoulant sur le cône alluvial) montre que certains secteurs dépourvus d'enjeux humains locaux peuvent faire l'objet

d'interventions de restauration morpho-sédimentaire, en particulier de suppression de protections de berges et d'arasement de seuils, sans risque particulier. Ce type d'action pourrait être étendu plus en aval et, surtout, plus en amont, de façon relativement aisée. Ce type de démarche adaptative devrait évidemment aussi être déployé sur la Moselotte et la Vologne.

Chapitre 6

Résumé étendu

Introduction/Contexte

La dynamique morpho-sédimentaire dans le massif vosgien reste aujourd'hui mal connue. Cette lacune touche autant les flux sédimentaires aux échelles temporelles actuelle et sub-actuelle que les trajectoires temporelles longues des stocks sédimentaires des fonds de vallées postérieurs au dernier maximum glaciaire, la temporalité longue étant fonctionnellement dépendante de la temporalité courte (Cordier et al., 2017). Or, la compréhension de la dynamique morpho-sédimentaire actuelle d'un système fluvial est un pré-requis indispensable à l'élaboration de stratégies de gestion, voire de restauration, efficaces et durables, des systèmes fluviaux (Maaß et al., 2021). A l'échelle du Massif des Vosges, ce travail de thèse, d'ordre à la fois fondamental et appliqué, met donc l'accent sur l'articulation des processus morpho-sédimentaires à des échelles temporelles emboîtées, depuis la production sédimentaire pluri-millénaire aux échelles spatiales du massif et des bassins versants, jusqu'aux ajustements géomorphologiques pluri-séculaires de tronçons de lits fluviaux.

Le Massif des Vosges présente trois caractéristiques majeures qui en font un objet particulièrement propice à l'étude et à la comparaison de la dynamique morpho-sédimentaire de ses systèmes fluviaux. Premièrement, l'origine géologique de sa formation explique une forte variabilité Nord-Sud de ses caractéristiques lithologiques, morphométriques et climatiques (Fig. 1.8; 1.9C). Deuxièmement, il présente la rare particularité de n'avoir été englacé que partiellement, c'est-à-dire uniquement dans sa partie Sud (Fig. 1.9A; Flageollet, 2002). Troisièmement, le Massif des Vosges est le plus densément peuplé de France (Parmentier, 2019), ce qui constitue à la fois un facteur de contrôle potentiel de la dynamique morpho-sédimentaire moderne, dont il convient de déterminer l'ampleur, et un enjeu en termes de gestion des hydrosystèmes (Fig. 1.15). Ce travail de thèse vise donc à caractériser le fonctionnement morpho-sédimentaire du Massif des Vosges, en cherchant à répondre à la problématique

suivante : Dans quelle mesure le fonctionnement actuel et sub-actuel des cours d'eau est-il impacté par les héritages (fluvio-)glaciaires et anthropiques ?

Objectifs/Méthodes

Ce travail de thèse s'est décliné en trois grands objectifs. Chaque objectif étant associé à une échelle spatiale et temporelle différente (Tableau 1.1).

Le premier objectif est d'évaluer l'état d'équilibre sédimentaire du massif des Vosges et de mesurer les taux de dénudation des bassins versants vosgiens. L'échelle spatiale d'étude correspond au massif dans son entièreté (Fig. 1.12A). L'échelle temporelle d'étude correspond à l'Holocène. Les concentrations en nucléides cosmogéniques (^{10}Be et ^{26}Al) ont été mesurées dans des échantillons de sédiments prélevés dans le lit des 21 rivières drainant l'ensemble du massif des Vosges. Un échantillon par rivière a été prélevé, au débouché du massif. La mesure des concentrations en ^{10}Be - ^{26}Al a permis d'une part d'évaluer l'état d'équilibre cosmogénique de chaque bassin versant en analysant le ratio $^{26}\text{Al}/^{10}\text{Be}$, et d'autre part, de calculer les taux de dénudation pluri-millénaire dans les bassins à l'équilibre. En parallèle, la variabilité spatiale des principaux potentiels facteurs de contrôle de la dénudation a été évaluée à l'échelle du massif. À savoir, la lithologie, la morphométrie des bassins, les précipitations et les héritages glaciaires (extension des anciens glaciers et distribution des dépôts glaciaires).

Le deuxième objectif est de reconstituer l'évolution morphosédimentaire post-LGM d'un système fluvial fortement impacté par les glaciations quaternaires. L'échelle spatiale d'étude correspond au bassin versant de la Cleurie. L'échelle temporelle d'étude correspond à l'Holocène. Le bassin versant de la Cleurie (versant ouest du massif) a été sélectionné car il est situé au cœur de la zone autrefois englacée du Massif des Vosges (Fig. 1.11). Il présente ainsi des caractéristiques géomorphologiques singulières à l'échelle du massif. Tout d'abord, il concentre la plus forte proportion de dépôts glaciaires et fluvio-glaciaires parmi tous les bassins versants vosgiens autrefois englacés (>40 % de sa surface, (Jautzy et al., 2024b)). Ensuite, des sondages géologiques (BRGM) indiquent des épaisseurs de sédiments en fonds de vallées dépassant 50 m dans la partie supérieure du bassin. Enfin, ce bassin versant de petite taille (<80 km²) englobe une très large gamme de morphologies et de dépôts associés, incluant des environnements glaciaires, alluviaux, fluvio-lacustres et marécageux. Sur base des travaux géomorphologiques précurseurs entrepris entre le milieu du XIX^e siècle (Hogard, 1840, 1842) et la fin du XX^e siècle (Flageollet et al., 1971; Flageollet, 1988), de prospections de terrain et d'un MNT à haute résolution, une cartographie géomorphologique détaillée du fond de vallée a été réalisée. Le degré d'ajustement du réseau hydrographique a été évalué à partir de l'analyse (i) des profils en long du chenal principal et des principaux affluents, et (ii) de l'évolution longitudinale de la granulométrie de surface du fond du lit de la Cleurie,

dont 1400 particules ont été mesurées (Wolman, 1954). Enfin, deux profils sédimentaires clés ont été étudiés, à partir notamment de mesures granulométriques et de datations radiocarbone. En parallèle, deux profils sédimentaires supplémentaires ont été étudiés dans le fond de vallée du bassin de la Fecht (versant est du massif). Ils correspondent respectivement à une berge active et une terrasse alluviale.

Le troisième objectif est de reconstituer la trajectoire temporelle sub-actuelle, sur les deux derniers siècles, des lits fluviaux et de leurs fonds de vallées et d'évaluer la part respective des divers facteurs de contrôle, notamment anthropiques. Les rivières concernées sont la Fecht, la Moselotte et la Vologne (Fig. 1.12B). L'étude repose sur des tronçons élémentaires de 500 m de l'amont vers l'aval des rivières. Sur la base du référentiel national des Obstacles à l'Écoulement (ROE, 2019), d'un MNT à haute résolution et de la prospection approfondie des archives documentaires de l'administration des Ponts et Chaussées, une base de données originale a été construite, recensant l'intégralité des seuils, digues et ponts présents dans le lit et le fond de vallée des trois cours d'eau. Les archives documentaires ont permis d'associer des périodes de (dé)construction de certains seuils. L'évolution planimétrique diachronique des lits actifs a été reconstruite à partir de la carte d'Etat-Major (1832) et de deux orthophotographies (1951, 2018). L'influence potentielle des facteurs de contrôle géomorphologiques a été évaluée en mesurant l'évolution longitudinale de la largeur du fond de vallée, de la présence de dépôts glaciaires et de la pente du talweg. Les patrons longitudinaux de la granulométrie de surface du fond des lits actifs (5000 particules) ont également été mis en évidence. Enfin, à l'échelle des bassins versants, l'évolution de l'occupation du sol entre le début du XIX^e et le début du XXI^e siècle a été mesurée à partir de la digitalisation de la carte d'Etat-Major et des données actuelles (OCSGE, Région Grand Est).

Résultats

Objectif 1 : Équilibre cosmogénique et dénudation pluri-millénaire du Massif des Vosges

Les résultats montrent que près de la moitié des bassins versants étudiés (10/22) sont en déséquilibre ($^{26}\text{Al}/^{10}\text{Be}$ compris entre 1,4 et 5,2; Fig. 2.5). La majorité de ces bassins sont situés dans la partie sud du massif, anciennement englacée. Une relation négative significative entre le ratio $^{26}\text{Al}/^{10}\text{Be}$ et la surface des dépôts glaciaires démontre une histoire d'exposition complexe, pouvant être expliquée par des héritages cosmogéniques dans le socle rocheux et/ou des épisodes de stockage/remaniement dans les fonds de vallées.

Pour les bassins versants à l'équilibre, aucune relation significative n'a pu être démontrée entre les taux de dénudation mesurés (compris entre 34 ± 1 et 66 ± 2 mm/ka pour le ^{10}Be)

et les facteurs de contrôle morphométriques, climatiques et lithologiques. Plus généralement, ces résultats soulignent l'importance de mesurer systématiquement la concentration en ^{26}Al , conjointement au ^{10}Be , afin de tester l'état d'équilibre des bassins versants, préalablement au calcul des taux de dénudation.

Objectif 2 : Ajustement post-LGM d'un système fluvial fortement impacté par les glaciations (Cleurie)

La cartographie géomorphologique permet de mettre en évidence une terrasse alluviale discontinue et peu développée (3,3 m au-dessus du talweg) et de confirmer l'existence de dépôts caractérisés par une surface plane et localisée à différents niveaux d'altitude distincts. Ces dépôts pourraient coïncider avec des niveaux de vidange successifs d'un paléolac, mis en place lors de la convergence de trois langues glaciaires (Flageollet, 1988). Les datations radiocarbones réalisées dans les profils sédimentaires en fond de vallée permettent de suggérer une vidange totale du paléolac à la sortie du LGM. Enfin, l'analyse des profils en longs du réseau hydrographique du bassin de la Cleurie ainsi que l'évolution longitudinale de la granulométrie de la surface du fond du lit (croissante vers l'aval), montrent un fort contrôle du fonctionnement hydro-géomorphologique de la Cleurie par les héritages glaciaires (Fig. 3.6).

Objectif 3 : Impact de l'anthropisation sur la trajectoire géomorphologique de la Fecht, Moselotte et Vologne

Les résultats permettent une première quantification de la pression anthropique industrielle sur les lits et fonds de vallées des trois cours d'eau. Un total de 125, 59 et 55 seuils sont actuellement présents dans le lit de la Fecht, Moselotte et Vologne, respectivement. Une période de construction a pu leur être attribuée à 10, 32, et 41 %, respectivement. La majorité d'entre eux ont été construits au milieu du XIX^e siècle pour la Fecht, tandis que l'aménagement des cours d'eau lorrains (Moselotte et Vologne) semble avoir été initié un peu plus tardivement tout en s'étendant sur des périodes plus longues. À l'inverse, un total de 4, 13 et 21 seuils ont été démantelés ou remobilisés par les crues depuis le XIX^e siècle, respectivement. Enfin, les facteurs de contrôle géomorphologique structurent, longitudinalement, le contrôle des aménagements hydrauliques sur les trajectoires évolutives des lits fluviaux. À l'échelle de certains tronçons fluviaux à fonds de vallées larges, une corrélation spatio-temporelle est observée entre le niveau d'anthropisation et l'intensité de la métamorphose fluviale observée. Ce dernier type d'évolution est particulièrement marquant pour la Fecht, dont le style fluvial est passé d'anabranches et de tresses, à un chenal unique quasi-intégralement figé, en moins d'un siècle (Fig. 4.13A).

Conclusions

Les résultats de ce travail de thèse mettent en évidence une influence significative des héritages glaciaires et anthropiques sur la dynamique morpho-sédimentaire des rivières vosgiennes. A l'échelle du Massif des Vosges, il a été démontré que près de la moitié des bassins versants sont en déséquilibre sédimentaire, plus ou moins prononcé. Le degré de déséquilibre est directement lié au degré d'englacement et de présence de dépôts (fluvio-)glaciaires dans les bassins versants. Pour les bassins anciennement englacés, les ratios $^{26}\text{Al}/^{10}\text{Be}$ indiquent des temps d'enfouissement apparents compris entre 0,5 et 3,4 Ma, ce qui suggère des temps de transit et de stockage de sédiments dans les fonds de vallées dépassant très largement l'Holocène. Concernant le bassin versant de la Cleurie, le temps d'enfouissement apparent calculé à partir du ratio $^{26}\text{Al}/^{10}\text{Be}$ est de 3,4 Ma, ce qui correspond à la valeur maximum mesurée à l'échelle du Massif des Vosges. Ce bassin constitue ainsi un exemple extrême de déséquilibre sédimentaire. Au-delà de la reconstitution chronologique partielle de la déglaciation du bassin versant proposée dans ce travail, la variabilité des profils en long et de la granulométrie de surface du lit de la Cleurie illustrent parfaitement le contrôle des héritages glaciaires sur la capacité (ou plutôt l'incapacité) d'un système fluvial à se ré-équilibrer, donc sur les processus hydro-géomorphologiques actuels et sub-actuels. Ainsi, il semble raisonnable de suggérer que la dynamique morpho-sédimentaire des autres bassins versants vosgiens anciennement englacés soit également significativement impactée.

L'étude détaillée de la trajectoire temporelle pluri-séculaire des fonds de vallées et lits fluviaux de la Fecht, Moselotte et Vologne montre que les héritages glaciaires sont effectivement un des facteurs de contrôle de la dynamique morpho-sédimentaire des cours d'eau. En revanche, sur cette temporalité, la pression anthropique devient un facteur au moins aussi important dans le contrôle des ajustements fluviaux. Il convient donc de considérer tant les héritages (fluvio-)glaciaires qu'anthropiques pour comprendre le fonctionnement contemporain des rivières étudiées. Enfin, afin d'évaluer la capacité de ces trois cours d'eau à remobiliser du matériel (fluvio-)glaciaire présent dans le fond de vallée, un total de 1350 traceurs RFID ont été mis en place sur 13 sites, couvrant l'ensemble du linéaire des trois cours d'eau étudiés. Un troisième levé des traceurs, prévu d'ici 2025, devrait apporter de nouveaux éclairages. Pour conclure, un diagnostic géomorphologique fonctionnel de la dynamique morpho-sédimentaire des rivières vosgiennes a ainsi pu être dressé, à travers (i) un portrait global de la production sédimentaire vosgienne et de l'état d'équilibre du massif, (ii) une évaluation de l'impact des héritages glaciaires sur l'ajustement post-LGM du bassin versant le plus fortement impacté par les glaciations et (iii) une évaluation de la part respective des facteurs de contrôle anthropiques et des héritages morpho-glaciaires, sur la trajectoire évolutive pluri-séculaire de trois cours d'eau.

Ces nouvelles connaissances régionales constituent finalement un ensemble d'éléments innovants d'appui à la décision pour la gestion, voire la restauration fonctionnelle des systèmes fluviaux vosgiens. Elles doivent permettre de mieux anticiper les résultats de projets de restauration hydro-géomorphologiques, de mieux prioriser les secteurs et les actions à entreprendre, tout en renforçant la résilience au changement climatique et en maîtrisant les risques.

Annexe A

Supplementary material for Chapter 2

A.1 Supplementary text

A.1.1 Data and methodology

A.1.1.1 Lithological classification

A widely-used approach for assessing the impact of lithology on denudation rates consists of clustering catchments into subsets based on their dominant lithology (Mandal et al., 2015; Olen et al., 2016; Peifer et al., 2021). This is generally achieved using expert opinion or an arbitrarily set of relative value (e.g., >50% area). The underlying idea is that if lithology exerts a primary control on the relationship between denudation rates and topography or precipitation, one can expect that catchments with similar topographic/climatic characteristics but different lithologies would display significant differences in denudation rates. Given the lithological complexity observed in the VM, we did not rely on an arbitrarily value to cluster the catchments but rather chose to apply hierarchical clustering analysis (HCA) on our dataset of 22 catchments described with the BRGM reclassified lithological variables, i.e. marno-calcareous, metamorphic, plutonic, sandstone, volcanic, alluvial, colluvial, glacial/fluvioglacial. This clustering technique attempts to find groups of observations within a dataset such that the observations in different groups are different from each other while the observations within each group are more similar to each other, according to the describing variables. HCA also determines the optimal number of groups. In our case, for each catchment, the observations correspond to the respective proportions of the five lithologies and the three types of Quaternary deposits extracted from the BRGM map. We used the FactomineR package with the Ward's minimum variance method (Ward, 1963). The resulting clustering procedure led to four classes of catchments that we named according to their main

respective lithological characteristics (i) sedimentary (ii) metamorphic (iii) plutonic and (iv) volcanic.

A.1.1.2 Sample preparation and AMS measurements

Quartz grains were isolated using first magnetic separation and then chemical purification with successive 1/3 HCl + 2/3 H₂SiF₆ baths. Purified quartz fractions were then etched three times in concentrated HF (48%) solutions to finally dissolve ~30% of the total weight and thus eliminate adsorbed meteoric ¹⁰Be. Purified quartz was spiked with ~150 mg of the ASTER ⁹Be carrier solution (3025 ± 9 ppm, Merchel et al., 2008) before total digestion in concentrated HF (48%). The solutions were then evaporated to dryness and precipitates were dissolved in 1M HNO₃. Natural concentrations of ²⁷Al were measured from an aliquot of the obtained solution by ICP-OES on a Thermo Scientific ICAP6500. Beryllium and aluminium were separated from the remaining solution by successive anionic and cationic resin extractions and precipitations. The resulting beryllium and aluminium precipitates were oxidized to BeO and Al₂O₃ which then mixed with Nb and Ag powders, respectively, and finally loaded into targets for measurements of ¹⁰Be/⁹Be and ²⁶Al/²⁷Al ratios at the French AMS National Facility, ASTER, located at CEREGE in Aix-en-Provence. Beryllium data were calibrated directly against the STD11 standard (Braucher et al., 2015) with a ¹⁰Be/⁹Be ratio of $(1.191 \pm 0.013) \times 10^{-11}$. Aluminium measurements were performed against an in-house standard called SM-Al-11 with ²⁶Al/²⁷Al = $(7.401 \pm 0.064) \times 10^{-12}$ which has been cross-calibrated against the primary standards certified by a round-robin exercise (Merchel and Bremser, 2004). Analytical uncertainties (reported as 1σ) include uncertainties associated with AMS counting statistics, AMS external error (0.5% for ¹⁰Be), chemical blank measurement, and regarding ²⁶Al, ²⁷Al measurements. A few studies corrected the production rates for quartz content by assigning relative abundance of quartz for each lithology (e.g. Safran et al. (2005)). Here, beyond difficulties linked to the large variety of lithologies constituting the VM, the 1:50,000 vectorised geological map indicates that ~95% of these are quartz-bearing rock formations that are thus homogeneously distributed within the topography. Such correction was thus not performed as it would only fractionally affect the results.

A.1.1.3 Re-calculation of denudation rates in the Black Forest Mountains (Germany)

The published ¹⁰Be-based denudation rates of the Black Forest Mountains (Morel et al., 2003; Meyer et al., 2010; Glotzbach et al., 2013) have been recomputed to allow a correct comparison with the denudation rates of the VM published in this study. We used CosmoCalc version 3.1 (Vermeesch, 2007), based on the time-independent Lal (1991)/Stone (2000) scaling model. We downloaded the ¹⁰Be concentrations of the 50 samples existing in the Black Forest Mountains,

from the OCTOPUS database (Codilean et al., 2018). The latter provides CRN concentrations renormalized to the 07KNSTD standard, which is equivalent to the STD-11 standard used in this study. We used the GIS shapefiles of the 50 catchments to compute the latitude of their centroid. For the sea-level-high-latitude (SLHL) reference production rates of ^{10}Be , we used the default value implemented in CosmoCalc version 3.1 which is 4.3 at/g/a. Muonic productions was based on the scheme of Braucher et al. (2011). We assumed a standard bedrock density of 2.7 g/cm^3 and did not apply any production correction for topographic shielding as reported by DiBiase (2018) nor for snow shielding.

A.1.2 Results

A.1.2.1 Lithology

We describe here the lithological specificities of each group of catchments resulting from the hierarchical clustering analysis of lithological units (see Section 2.3.1; Fig. 2.3). Sedimentary catchments (*s*; catchments #1, 9, 16, 15, 10, 21, 22, 11) are located in the northern part of the VM, on both sides of the drainage divide, and display a very homogeneous sedimentary cover mainly composed of sandstone and marno-calcareous. Their cumulated proportions range from 41 to 93%. Quaternary deposits are mainly composed of alluvial and colluvial deposits, ranging from 7 to 25%. In the southern part of the VM, studied catchments correspond to metamorphic (*m*), plutonic (*p*), and volcanic (*v*) classifications, that all display much more variability in exposed lithologies than the sedimentary catchments. Metamorphic catchments (*m*, catchments #6, 8, 5, 7, 18) are all located on the eastern side of the VM and display at least 43% of metamorphic bedrock, complemented with plutonic, volcanic, and sandstone proportions. Quaternary deposits are of various types and proportions, ranging from 7 to 19%. Plutonic catchments (*p*; catchments #19, 2, 12, 13, 20, 4) are located on both sides of the VM and display at least 40% of plutonic bedrock, except for catchment 19 that displays 27% of plutonic bedrock. A minimum of 40% of the Quaternary deposits in the plutonic catchments are composed of glacial and fluvio-glacial deposits. Volcanic catchments (*v*; catchments #14, 17, 3) are located at the southernmost part of the VM. They display at least 35% of volcanic bedrock, complemented with plutonic, metamorphic, and sandstone proportions. As for the plutonic catchments, a significant proportion (>30%) of the Quaternary deposits are composed of glacial and fluvio-glacial deposits.

A.2 Supplementary figures and tables

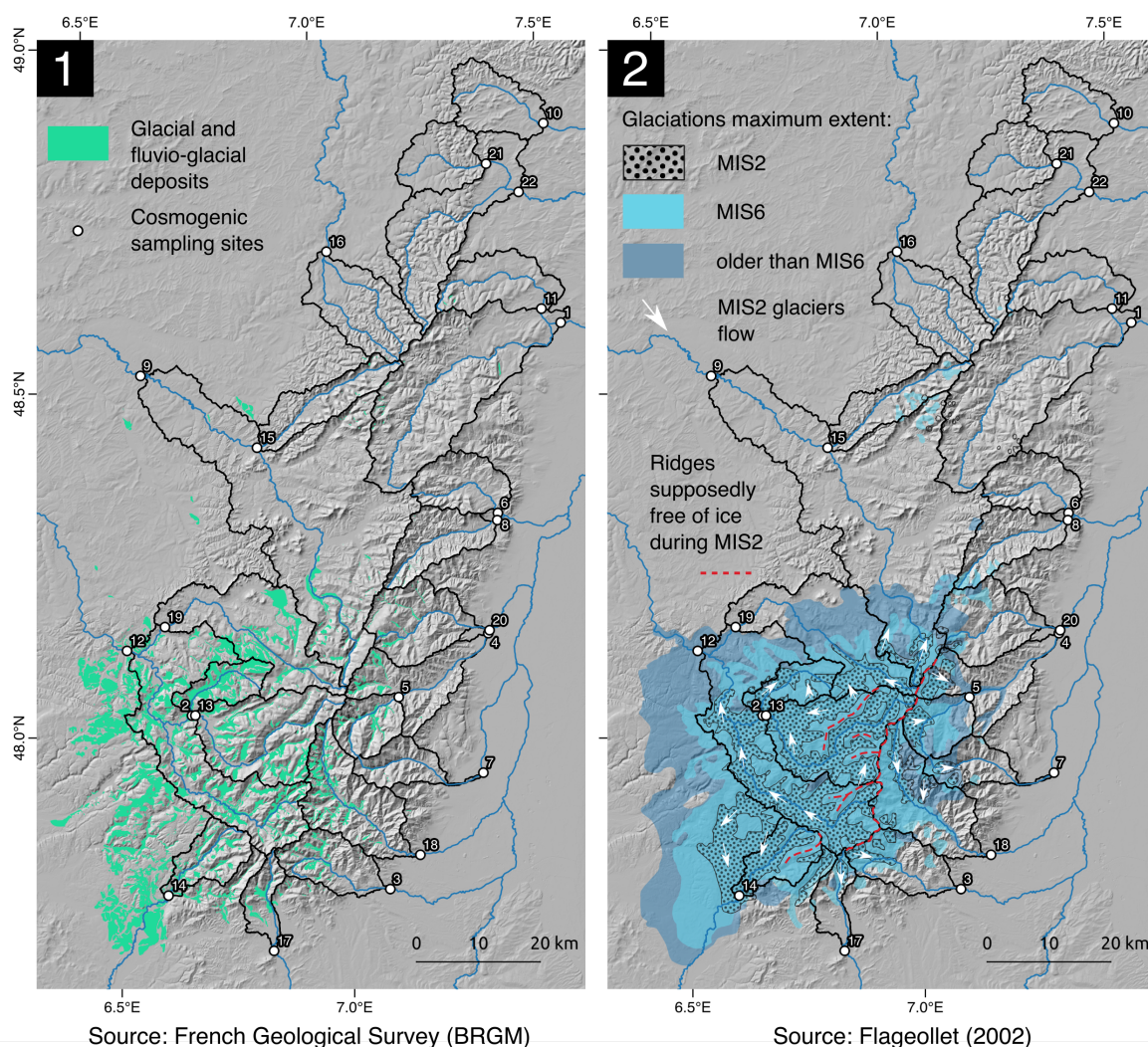


FIGURE A.1: Spatial data used in this study to estimate the coverage of glacial-related deposits and the former glaciations extents. 1: The coverage of glacial and fluvio-glacial deposits is derived from the BRGM geological map (shapefiles available online: <https://infoterre.brgm.fr/formulaire/telechargement-cartes-geologiques-departementales-150-000-bd-charm-50>). 2: The maximum extents of the supposedly two last (MIS2 and MIS6) and the 'pre-MIS6' glaciations corresponds to the ones mapped by Flageollet (2002). We coregistrated the maps and digitised the glaciations extents.

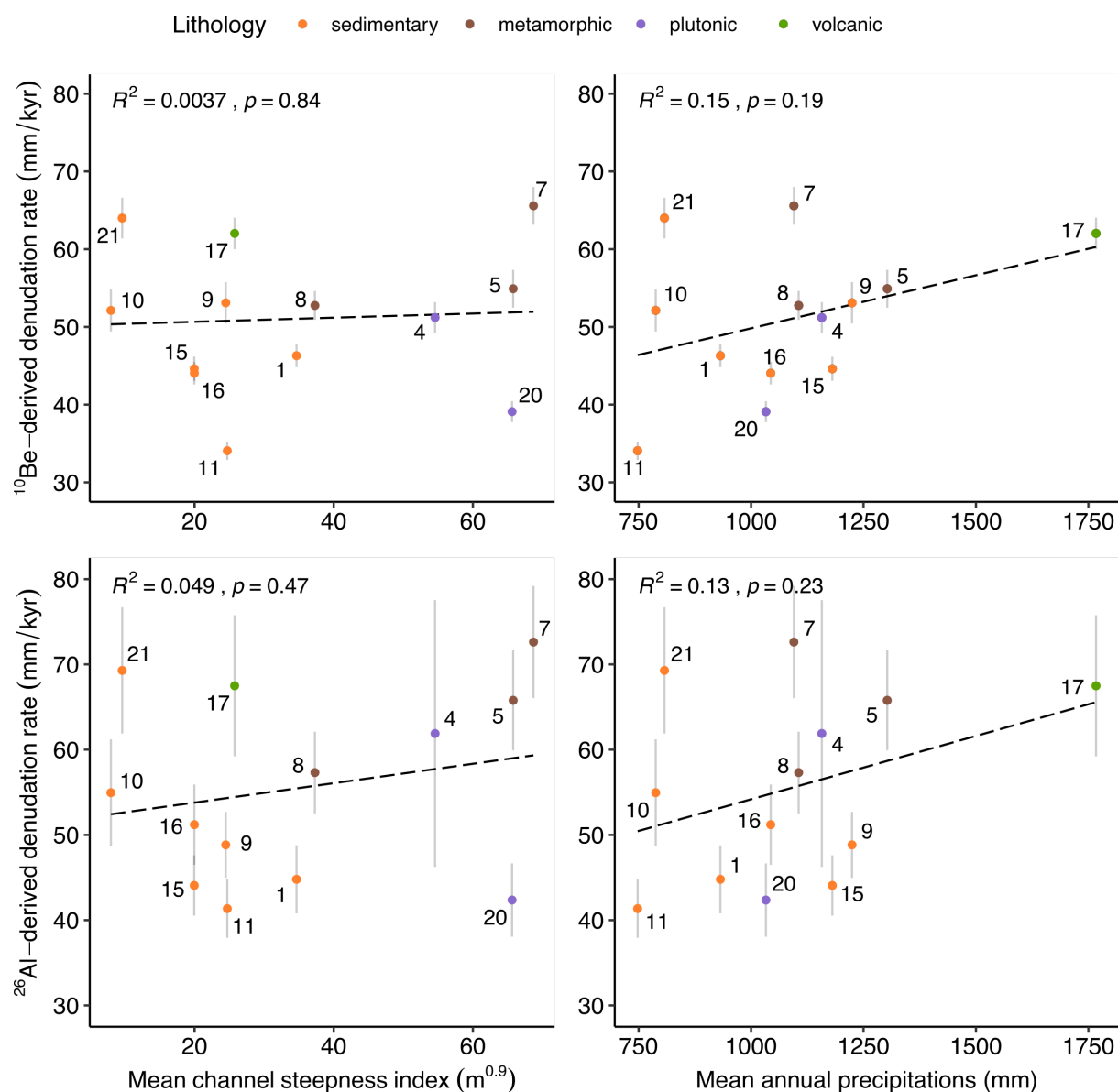


FIGURE A.2: ^{10}Be and ^{26}Al -derived CWD rates vs. mean channel steepness index and mean annual precipitations, excluding the *non-steady-state* catchments. No significant relation between denudation rates and potential controlling factors can be observed.

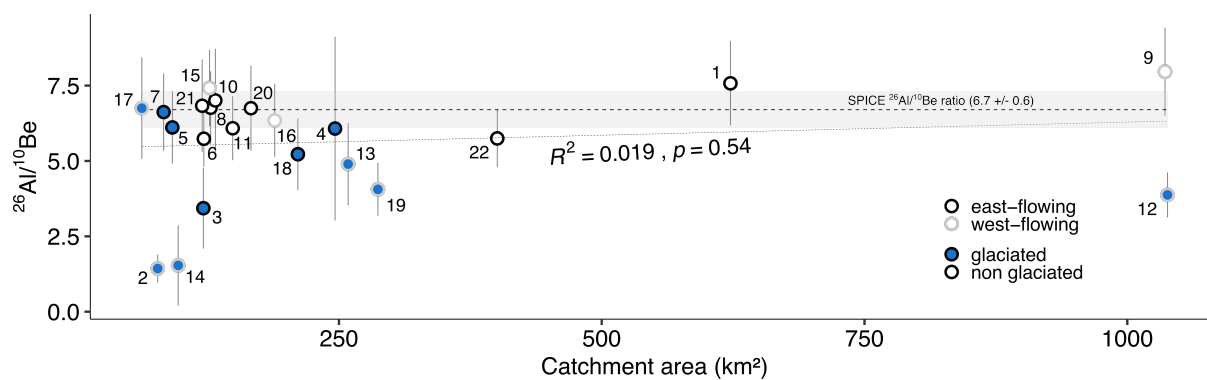


FIGURE A.3: Relationships between $^{26}\text{Al}/^{10}\text{Be}$ ratios and catchment area (km^2). The vertical error bars correspond to the 1σ uncertainty in the $^{26}\text{Al}/^{10}\text{Be}$ ratio. Blue points highlight the glaciated catchments (i.e., >10% of ice cover during MIS2). The horizontal dashed line corresponds to the surface production ratio recently determined by the SPICE project (6.7 ± 0.6 ; Fenton et al. (2022)). We did not use the surface production ratio from Cosmocalc (Vermeesch, 2007) as it does not provide any uncertainty.

TABLE A.1: Correlation coefficients from linear regressions between CWD rates and potential controlling factors. Correlation coefficients are computed from CWD rates inferred from both nuclides. We distinguished the whole dataset from the steady-state catchments and the eastern from the western catchments. No significant relation is observed when excluding the non-steady-state catchments. *: p-value<0.05; **: p-value<0.01

		Area	Mean elevation	Mean slope	Mean k_{SH}	Mean annual precipitations
All catchments (n=22)	¹⁰ Be-derived	all (n=22)	0.46*	0.37	0.20	0.47*
	CWD rates	eastern (n=13)	0.36	0.44	0.08	0.63*
		western (n=9)	0.79*	0.30	0.88**	0.42
	²⁶ Al-derived	all (n=22)	0.35	0.05	0.13	0.49*
	CWD rates	eastern (n=13)	0.33	0.39	0.04	0.67*
		western (n=9)	0.48	0.06	0.7*	0.28
Steady catchments (n=13)	¹⁰ Be-derived	all (n=13)	0.21	0.20	0.06	0.39
	CWD rates	eastern (n=9)	0.13	0.26	0.03	0.27
		western (n=4)	NA	NA	NA	NA
	²⁶ Al-derived	all (n=13)	0.34	0.33	0.22	0.36
	CWD rates	eastern (n=9)	0.28	0.38	0.16	0.41
		western (n=4)	NA	NA	NA	NA

TABLE A.2: Cosmogenic nuclide data from ^{10}Be and ^{26}Al measurements in quartz from stream sediment.

Sample name	Sample ID	Latitude (°)	Longitude (°)	Mass quartz (g)	^9Be spike (at)	Corrected ^{10}Be concentration (at/g)	1σ ^{10}Be uncertainty (at/g)	^{10}Be -derived denudation rate (mm/ka)	1σ uncertainty (mm/ka)	Integration time (ka)	Corrected ^{26}Al concentration (at/g)	1σ ^{26}Al uncertainty (at/g)	^{26}Al -derived denudation rate (mm/ka)	1σ uncertainty (mm/ka)	Integration time (ka)	$^{26}\text{Al}/^{10}\text{Be}$	Uncertainty $^{26}\text{Al}/^{10}\text{Be}$
br	1	48.57	7.53	21.90	3.15×10^{19}	9.77×10^4	3029.90	46.30	1.47	12.96	7.40×10^5	63845.93	44.79	3.99	13.40	7.57	0.69
cl	2	48.02	6.68	18.79	3.13×10^{19}	7.73×10^4	4560.41	69.60	4.17	8.62	1.11×10^5	16231.99	362.56	53.42	1.65	1.43	0.23
do	3	47.76	7.09	20.81	3.17×10^{19}	6.06×10^4	2437.68	87.66	3.57	6.84	2.08×10^5	39470.36	189.16	36.22	3.17	3.43	0.67
fe-01	4	48.12	7.33	18.82	3.16×10^{19}	1.05×10^5	4008.18	51.20	1.99	11.72	6.38×10^5	157415.63	61.89	15.64	9.69	6.08	1.52
fe-02	5	48.04	7.13	19.93	3.15×10^{19}	1.10×10^5	4792.44	54.91	2.43	10.93	6.73×10^5	58789.49	65.78	5.86	9.12	6.11	0.60
gi	6	48.30	7.36	19.12	3.15×10^{19}	1.22×10^5	3835.36	36.45	1.18	16.46	7.02×10^5	51832.61	46.93	3.58	12.79	5.74	0.46
la	7	47.92	7.30	21.15	3.18×10^{19}	8.53×10^4	3116.54	65.58	2.43	9.15	5.65×10^5	50184.80	72.62	6.58	8.26	6.62	0.64
li	8	48.28	7.36	21.88	3.14×10^{19}	9.11×10^4	3136.86	52.76	1.85	11.37	6.16×10^5	50188.16	57.31	4.79	10.47	6.76	0.60
me	9	48.52	6.60	19.29	3.15×10^{19}	8.72×10^4	4284.49	53.10	2.66	11.30	6.95×10^5	53385.48	48.83	3.87	12.29	7.96	0.73
mod	10	48.86	7.51	19.04	3.15×10^{19}	7.35×10^4	3712.70	52.12	2.70	11.51	5.15×10^5	56897.61	54.95	6.26	10.92	7.01	0.85
mog	11	48.59	7.48	21.86	3.15×10^{19}	1.20×10^5	4012.90	34.07	1.18	17.61	7.30×10^5	58208.73	41.35	3.42	14.51	6.09	0.53
mos	12	48.12	6.54	21.00	3.15×10^{19}	9.82×10^4	3438.84	53.67	1.91	11.18	3.81×10^5	33796.00	102.69	9.26	5.84	3.87	0.37
mot	13	48.02	6.69	19.78	3.08×10^{19}	7.13×10^4	3632.91	80.69	4.16	7.44	3.49×10^5	44944.12	121.39	15.81	4.94	4.90	0.68
og	14	47.76	6.61	20.32	3.10×10^{19}	9.27×10^4	12517.22	54.78	7.53	10.95	1.42×10^5	58239.13	267.30	110.03	2.24	1.53	0.66
pl	15	48.41	6.85	21.30	3.17×10^{19}	1.02×10^5	3458.03	44.62	1.55	13.45	7.59×10^5	58911.24	44.07	3.54	13.61	7.42	0.63
sa	16	48.69	7.02	19.58	3.17×10^{19}	9.93×10^4	3197.59	44.05	1.45	13.62	6.30×10^5	56217.84	51.20	4.71	11.72	6.34	0.60
sav	17	47.68	6.83	21.86	3.16×10^{19}	8.28×10^4	2655.31	62.03	2.02	9.67	5.59×10^5	67073.16	67.49	8.28	8.89	6.75	0.84
th	18	47.80	7.16	21.45	3.13×10^{19}	6.82×10^4	2719.40	79.85	3.23	7.51	3.56×10^5	37625.07	112.67	12.07	5.33	5.22	0.59
vo	19	48.15	6.63	18.89	3.17×10^{19}	1.15×10^5	5680.60	44.81	2.27	13.39	4.66×10^5	44246.41	81.89	7.92	7.33	4.06	0.43
we	20	48.13	7.33	21.37	3.15×10^{19}	1.37×10^5	4602.60	39.10	1.34	15.35	9.25×10^5	91056.91	42.36	4.30	14.17	6.75	0.70
zis	21	48.80	7.38	21.32	3.19×10^{19}	6.17×10^4	2467.36	64.00	2.61	9.38	4.21×10^5	43909.67	69.29	7.40	8.66	6.83	0.76
zo	22	48.76	7.45	20.88	3.20×10^{19}	8.64×10^4	3370.19	46.99	1.88	12.77	4.97×10^5	35471.52	60.60	4.45	9.90	5.75	0.47

Annexe B

PickShift: A user-friendly Python tool to assess the surficial uncertainties associated with polygons extracted from historical planimetric data

This article is under review (minor revisions) in the journal **SoftwareX** (Jautzy et al., 2024a). It corresponds to the generalisation of the method developed by Jautzy et al. (2020) and its transformation into a Python tool.

Article abstract With the increasing use of GIS software's, historical planimetric data such as orthophotos and old maps represent key data sources to analyze spatio-temporal landscape evolution. However, geometric error inherent to these data are too often overlooked, possibly leading to confusing misinterpretation of measured surficial changes. The user-friendly Python tool 'PickShift', based on a Monte-Carlo approach, addresses this critical issue by quantifying the surficial uncertainty associated with any features digitized from historical planimetric data. This software provides a valuable framework for a more accurate assessment of landscape dynamics and associated uncertainties.

B.1 Motivation and significance

With the increasing use of geographic information system (GIS) software's, historical planimetric data (HPD), such as aerial photographs, orthophotos and historical maps, have become a major source of data to explore the spatio-temporal evolution of landscapes (Chiang et al.,

TABLE B.1: Code Metadata

Nr	Code metadata description	Information
C1	Current code version	v1.0
C2	Permanent link to code/repository used for this code version	https://github.com/r-wenger/Pickshift
C3	Permanent link to reproducible capsule	https://seafile.unistra.fr/d/a6a19d553ff24ee7859e/
C4	Legal code license	GPL-3.0
C5	Code versioning system used	'none'
C6	Software code languages, tools and services used	python
C7	Compilation requirements, operating environments and dependencies	/
C8	If available, link to developer documentation/manual	https://github.com/r-wenger/Pickshift
C9	Support email for questions	timothee.jautzy@live-cnrs.unistra.fr

2014; Herrault and Sheeren, 2022). This approach primarily consists in digitizing features of interest from diachronic, coregistrated planimetric data to analyze the decadal to centennial evolution of these features. It allows addressing various socio-environmental topics focusing on a wide range of extractable features, such as forests fragments (Herrault et al., 2013), ecotones (Chardon et al., 2022), river channels (Jautzy et al., 2022), valley glaciers (DeVisser and Fountain, 2015), urban objects (Xiao et al., 2006), and, more generally, land cover (Chen et al., 2019; Skaloš et al., 2011). It can eventually improve environmental management practices in contexts where quantifying impacts of human activities and climate change on socio-ecological systems helps to better understand landscape dynamics (Dufour et al., 2019). Moreover, this approach is simple and efficient since it only requires a GIS software and diachronic HPD, both becoming widely and freely available. For instance, the French National Geographic Institute (IGN) recently made available two series of aerial photographs (1950, 2018) and a historical map ('Etat-Major'; 19th century) covering the whole French territory in a coregistrated format (Lallemant et al., 2017).

However, the process of data coregistration inevitably distorts the raw data so that geometric errors are inherently associated with HPD (Jenny et al., 2007). Surprisingly, many studies dealing with the spatio-temporal evolution of landscapes or landforms from HPD either do not consider these geometric errors (Etter et al., 2008; Hooke and Yorke, 2010; Skaloš et al., 2011), or simply report the root mean square error (RMSE) calculated from the ground control points (GCP) used to coregistrate the initial data. This last approach assumes a uniform geometric error over the whole study area (DeVisser and Fountain, 2015; Jacob et al., 2016; Yousefi et al., 2019; Zanoni et al., 2008), although a spatial variability intrinsically affects these errors associated with HPD (Herrault et al., 2013). In fluvial contexts where the amplitude of

channel-planform changes might be too low to be detected (Lauer et al., 2017), Jautzy et al. (2020) and Leonard et al. (2020) demonstrated that considering the spatially variable error (SVE) allows to determine the (non-)significance of measured channel planform changes, such as lateral channel migration. Both developed a reproducible method to evaluate the impact of SVE on channel-planform changes measured from diachronic, coregistrated aerial photographs (Leonard et al., 2020) or orthophotos (Jautzy et al., 2020). Although the useful approach of both previous studies can be theoretically transposed to other features, they only focused on fluvial settings and none of them delivered a user-friendly tool in an open-source language.

We argue that these automatized approaches should be extended to any kind of features extractable from HPD to detect (i) the uncertainty of surfaces measured in different environmental settings or (ii) the (non-)significance of planimetric changes over time. We thus propose a python-based tool, called PickShift, to improve the generalization of methods allowing for the assessment of uncertainties associated with HPD (Donovan and Belmont, 2019). Based on a Monte-Carlo approach, it is an enhanced version of Jautzy et al.'s (2020) and assesses the uncertainties associated with the surfaces of features (i.e. polygons) digitized from any targeted HPD (i.e. aerial photographs, orthophotos and historical maps), compared to reference planimetric data (i.e. recent orthophoto). Its denomination merges the twofold action of 'pick'ing ground control points (GCP) and randomly 'shift'ing the polygons according to the SVE affecting the targeted data.

B.2 Software description

PickShift's main achievement allows computing the surficial uncertainty of polygons digitized from HPD. The script is primarily based on Jautzy et al.'s (2020) workflow but its simple design does not require specific Python skills to be used. The complete installation and operation guidelines are provided in the Github. PickShift consists of three files:

- **environment.xml**: It allows the operator to automatically configure and build an Anaconda environment that suits the appropriate operating system.
- **config.conf**: It contains the location of the required inputs and a few parameters that can be modified by the operator.
- **pickshift.py**: It is the main source code that has to be run once the Anaconda environment is created and the configuration file parameterized.

PickShift also requires three input files that must be prepared by the operator using any GIS software (e.g. QGIS, ArcGIS):

- a shapefile of the spatial extent of the studied area
- a shapefile of the features digitized from the HPD
- a set of independent GCPs

B.2.1 Software architecture

PickShift consists of two main steps and one optional step. It first interpolates the SVE over the study area and then assesses the surficial uncertainty of each polygon (Fig. B.1). The optional step allows simplifying the polygons' geometry, if needed.

B.2.1.1 Spatially-variable error (SVE) computing (step 1)

The software first measures the planimetric bias (i.e. Euclidean distance) for each pair of GCPs, by comparing the geographic coordinates picked from the reference planimetric data with those picked from the targeted data. Local planimetric biases are then interpolated over the spatial extent of the whole study area, using an inverse distance weighting method (Fig. B.1). The software then exports three rasters corresponding to the SVE in X, Y and XY. The resolution and the coordinates system of the SVE rasters are set by the operator. The mean SVE is then compared with the mean distance between the successive vertices of all polygons. If this ratio is >1 , the operator is warned that subsequent Monte-Carlo (MC) simulations may cause topological errors. The operator must choose between ignoring the warning to continue the analysis, or restarting it after setting Douglas-Peucker simplification parameters (optional step).

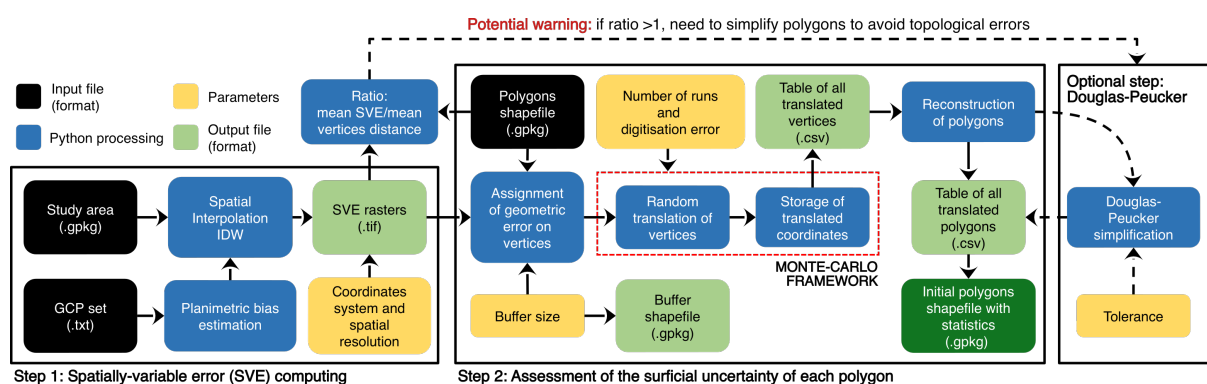


FIGURE B.1: PickShift workflow

B.2.1.2 Monte-Carlo simulations and assessment of the surficial uncertainty of each polygon (step 2)

For each polygon vertex, the values of all pixels from the SVE rasters within a buffer are selected. The size of the buffer is set by the operator. The normal distribution of SVE around each vertex is then calculated by averaging the mean local error and by calculating the standard deviation. Hence, for each MC run, a specific value of SVE in X and Y is randomly extracted from the respective normal distribution and used to shift each vertex from its original position (Fig. 1). The direction of the translation in X and Y also are randomly selected at each MC run with equal probability weights (i.e. 50% each). The coordinates of the translated vertices are stored for each run. The number of runs is set by the operator.

The simulation process can be represented by the following mathematical expressions:

$$\begin{aligned} X_{\text{changed}} &= X_{\text{original}} + |e_x| \times \beta_1 + e_d \times \beta_2 \\ Y_{\text{changed}} &= Y_{\text{original}} + |e_y| \times \beta_1 + e_d \times \beta_2 \end{aligned}$$

Here, $|e_x|$ and $|e_y|$ represent the absolute registration errors in the X and Y directions, respectively. The constant e_d represents the digitizing error, which is set by the operator. β_1 and β_2 are coefficients that determine the direction of the registration error ($|e_x|$ and $|e_y|$) and the digitizing error (e_d), and they are randomly selected during each run. Note that for each polygon, β_1 remains constant for each X_{changed} (i.e., constrained in only one direction). This constraint allows maintaining the directional sequence of successive vertices (Jautzy et al., 2020).

Once the Monte-Carlo simulations are completed, polygons are reconstructed (poly_MC.gpkg) from the stored coordinates (point_sim_MC.csv). If the operator chooses to simplify the reconstructed polygons, the Douglas-Peucker algorithm is applied according to the tolerance parameter (see part 2.1.3). The area of the reconstructed polygons is then computed, along with a series of statistics associated with each of the initial polygons. PickShift automatically computes the following statistics from the total range of measured values compiled from all MC runs:

- average area;
- standard deviation;
- minimum and maximum;
- first, second and third quartile;
- total and 95% uncertainty (Jautzy et al., 2020).

Finally, the software exports the final shapefile (`poly_MC.gpkg`), corresponding to the initial shapefile to which the previously calculated statistics are appended.

B.2.1.3 Douglas-Peucker simplification (optional step)

This optional step is required if the ratio between the mean SVE and the mean distance between vertices is > 1 (Fig. 1). In that case, Monte-Carlo simulations may cause topological errors, leading to fundamental incorrectness in the calculation of polygon surfaces. A practical solution consists in performing geometric simplification of the reconstructed polygons, after the Monte-Carlo simulations. PickShift implements the Douglas-Peucker algorithm Douglas and Peucker (1973), subdividing the initial line or polygon into smaller segments and joining their endpoints with straight lines. Then, it removes any points closer to the straight line than a specified tolerance threshold. If the aforementioned ratio is > 1 , PickShift automatically suggests a tolerance value to be set by the operator. The tolerance suggestion corresponds to twice the mean distance between vertices.

B.2.2 Technical recommendations

PickShift is designed to quantify the SVE of coregistrated data, from a set of GCPs selected by the operator. If the latter coregistrates the data, the set of GCPs used to interpolate the SVE must be independent from the one initially used to coregistrate the data Hughes et al. (2006); Jautzy et al. (2020). GCPs should be uniformly distributed over the whole study area and hard features (e.g. buildings, roads...) are preferred over soft ones (e.g. pathway intersections, trees...). A minimum of six to eight GCPs is recommended as, beyond this threshold, the quality of error characterization is not improved Hughes et al. (2006).

We recommend interpolating the SVE with a resolution equal to that of the targeted planimetric data. A digitizing error of at least one pixel (referring to the targeted data) is considered reasonable since the user is not able to see details below this value. Concerning the buffer radius used to extract SVE values around vertices, we suggest applying 10 times the resolution of the targeted planimetric data. Finally, if a Douglas-Peucker simplification is performed, we strongly recommend a visual inspection of the simplification results before any interpretation of the surficial uncertainty results. We suggest to apply first a tolerance value equal to twice the average distance between vertices and to gradually increase it as long as topological errors persist. One must keep in mind that any geometrical simplification necessarily results in a degradation of the initial data (see part 3).

B.3 Illustrative examples

Both examples presented here aim to illustrate the use of PickShift on two types of data widely used in planimetric studies: a coregistrated historical map and an orthophoto. Presenting contrasting characteristics in terms of spatial resolution, geometric error and topography, the specificities of their use in PickShift are highlighted accordingly. We also aim to assess the impact of polygons simplification using the Douglas-Peucker algorithm (part 2.1.3). The targeted planimetric data from the first dataset (Munster) is the 'Etat Major' map, produced between 1820 and 1856 and covering the whole French territory. It is provided by the IGN in a coregistrated format (EPSG:2154) and has a resolution of 4 m. The targeted planimetric data from the second dataset (Bruche) is an aerial photograph surveyed in 1950 and covering the whole French territory. It is provided by the IGN in an orthorectified format (EPSG:3948) and has a resolution of 0.50 m. For both datasets, the reference planimetric data used to create the GCP set is an orthophoto from 2015 provided by the IGN with a resolution of 0.20 cm.

PickShift was applied on both datasets (Munster and Bruche) to assess the surficial uncertainty of digitized polygons through 1000 Monte-Carlo simulations, with and without an additional Douglas-Peucker simplification. The same three types of polygons were digitized from both datasets: river channel, cropland and forest. The buffer was set to ten times the spatial resolution of the targeted planimetric data. The spatial resolution of the SVE rasters and the digitization error were set to the spatial resolution of the targeted planimetric data.

B.3.1 Munster dataset

This dataset is composed of sixteen polygons manually digitized from the 'État major' map (1820/1856) centered around the town of Munster along the Fecht valley (Haut-Rhin/France; Fig. B.2A). High topographic variations characterize this area (Table B.2). We selected eighteen GCPs to estimate the planimetric bias between the targeted ('Etat major' map) and the reference (2015 orthophoto) planimetric data (Fig. 2A). The mean planimetric bias in XY amounts to 34.38 m (± 27.80 m). The interpolated SVE ranges from 5.9 to 94.8 m (Fig. B.2B). Here, the ratio between the mean SVE and the mean distance between vertices is >1 (i.e., 3.2). Figure B.3A and B.3B shows the first ten translations of polygon #3 resulting from Monte-Carlo simulations without and with an additional Douglas-Peucker simplification, respectively. One must bear in mind, however, the incorrectness of calculating the area of a polygon whose geometry contains topological errors. Here, the Douglas-Peucker simplification is mandatory to remove these. We used a tolerance of 50 m. The total surficial uncertainty associated to the polygons without and with an additional Douglas-Peucker simplification ranges from 4.06 to 39.98 % and 7.44 to 62.80 %, respectively (Fig. B.4A).

TABLE B.2: Main characteristics of the two datasets

Dataset	Targeted planimetric data	Spatial resolution (m)	Projection system (EPSG)	Statistics	Number of GCP	Number of polygons	Mean planimetric bias XY (m)
Munster	État major map (1820/1856)	4.0	Lambert 93 (2154)	mean:480.4 std: 84.5 min: 339.7 max: 785.5	18	16	34.38 +/- 27.80
Bruche	Orthophoto (1950)	0.5	RGF93/CC48 (3948)	mean:143.8 std: 1.3 min: 139.1 max: 148.4	12	12	0.92 +/-0.45

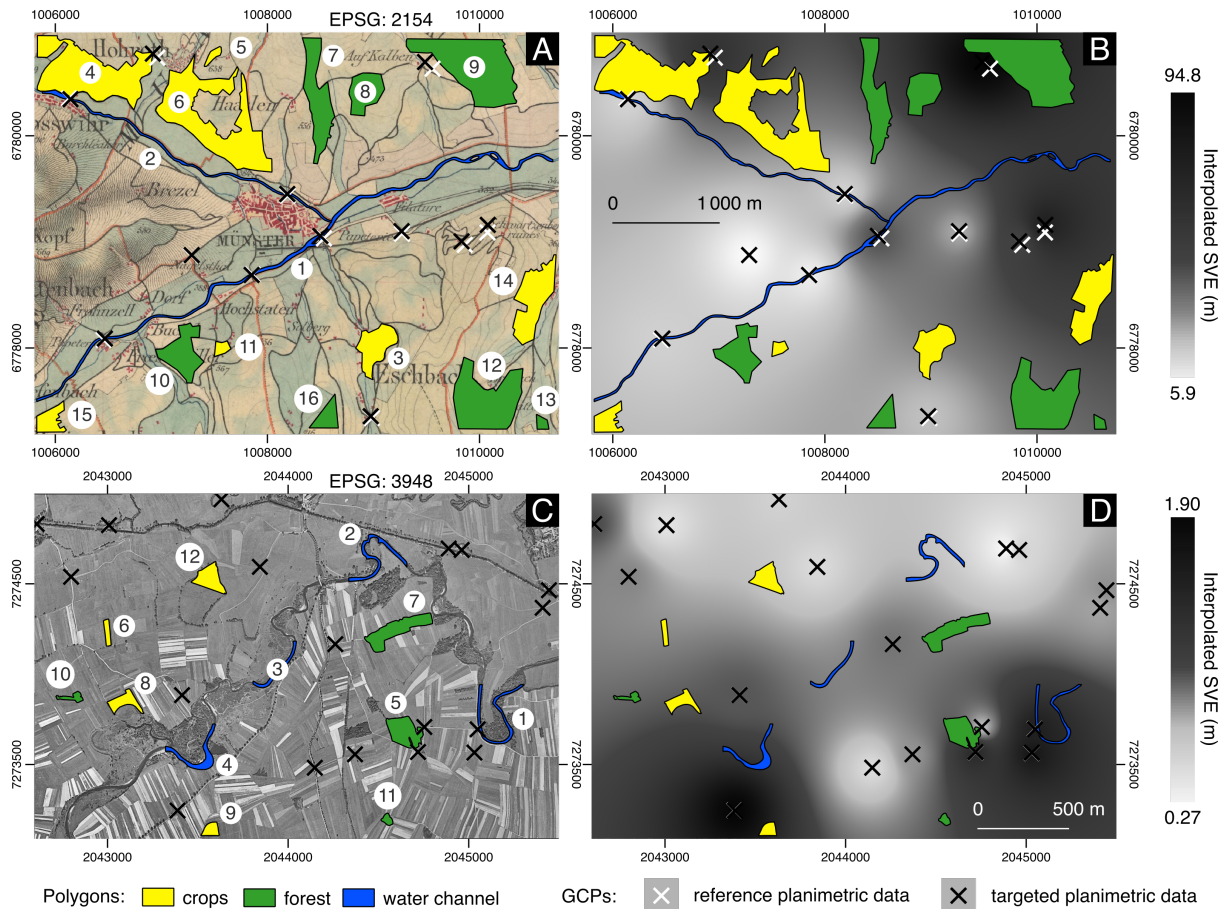


FIGURE B.2: Datasets used in the illustrative examples showing the three types of polygons digitized from the targeted planimetric data, GCP set and interpolated SVE for the Munster (A-B) and the Bruche dataset (C-D).

B.3.2 Bruche dataset

This dataset is composed of twelve polygons manually digitized from an orthophoto (1950) centered on the Bruche River's floodplain located five kilometers westward of Strasbourg city (Bas-Rhin/France; Fig. B.2C). Low topographic variations characterize this area (Table B.2). We selected twelve GCPs to estimate the planimetric bias between the targeted (1950 orthophoto) and the reference (2015 orthophoto) planimetric data (Fig. B.2C). The mean planimetric bias in XY amounts to 0.92 m (± 0.45 m). The interpolated SVE ranges from 0.27 to 1.90 m (Fig. B.2D). As the ratio between the mean SVE and the mean distance between vertices is < 1 (i.e., 0.05), the optional step (Douglas-Peucker) is not necessary but it was performed for the sake of comparison with the first dataset (Fig. B.3C-D). We used a tolerance of 1 m. The total surficial uncertainty associated with the polygons ranges from 1.28 to 4.80% and 1.18 to 5.04% without and with an additional Douglas-Peucker simplification, respectively (Fig. B.4B).

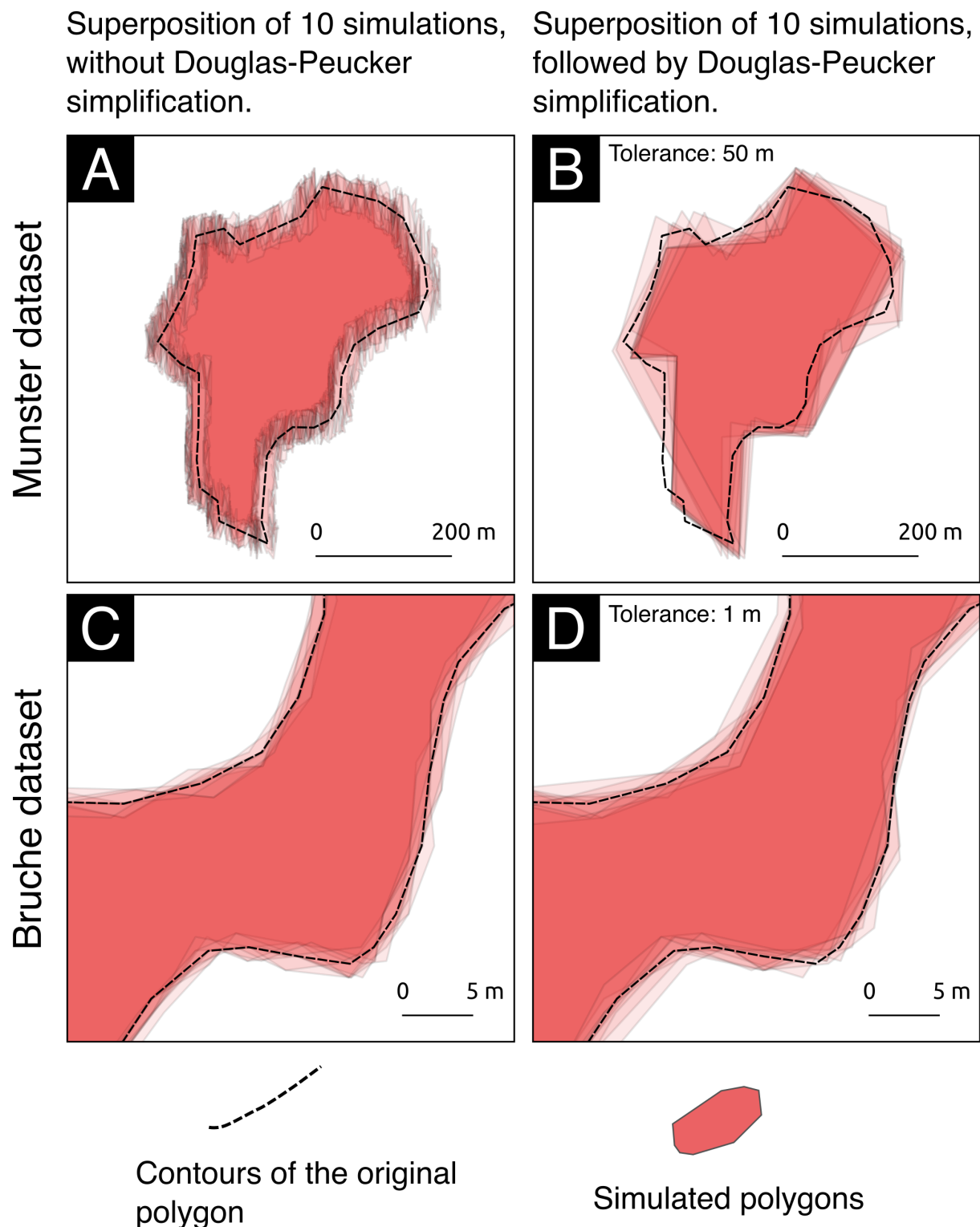


FIGURE B.3: Examples of ten simulations for polygon #3 of the Munster dataset, without (A) and with (B) Douglas-Peucker simplification and for polygon #2 of the Bruche dataset, without (C) and with (D) Douglas-Peucker simplification.

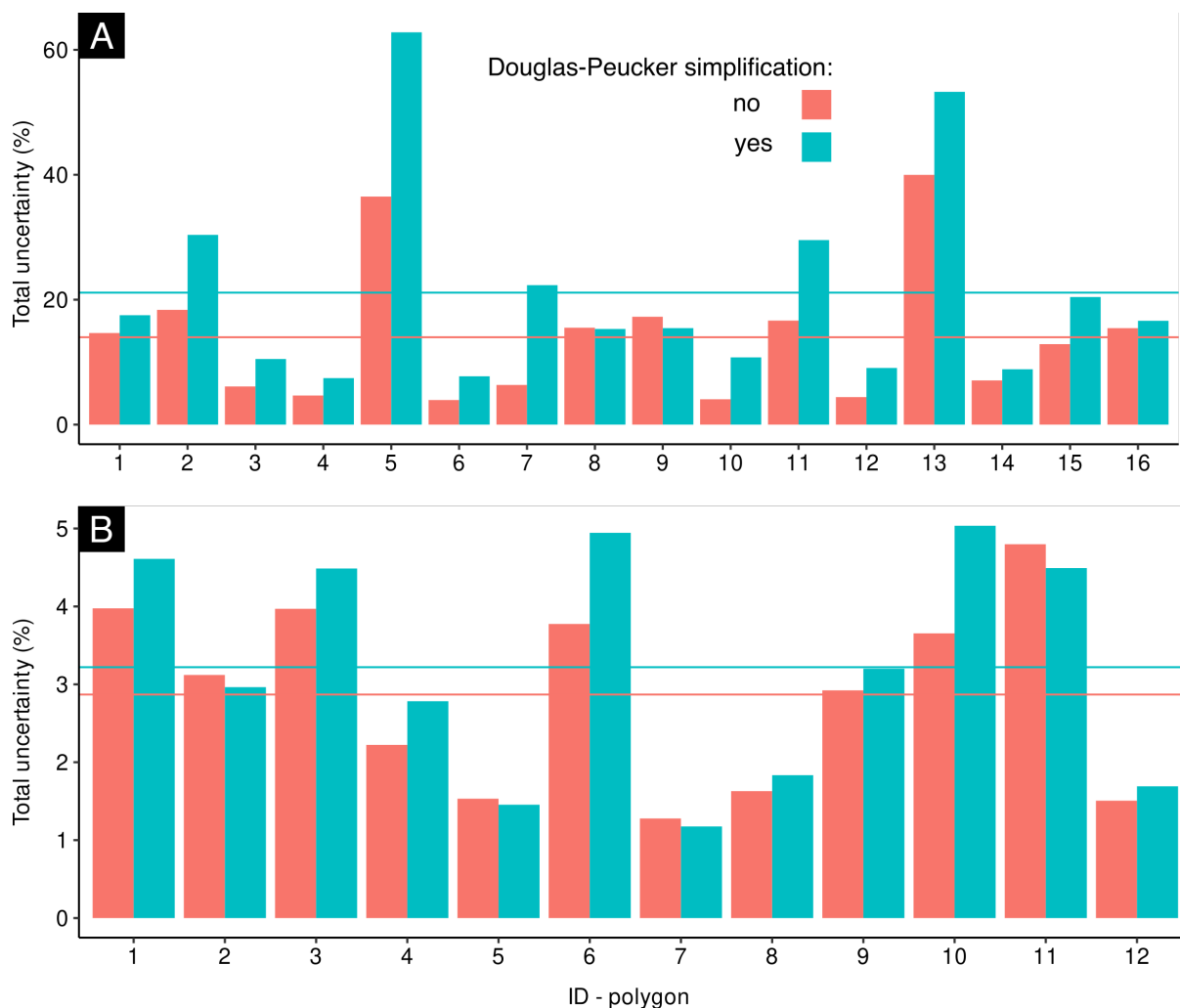


FIGURE B.4: Surficial uncertainty associated with digitized polygons from Munster (A) and Bruche (B) dataset, without and with applying the Douglas-Peucker simplification.

B.4 Impact

PickShift is a valuable tool enhancing the reliability of studies based on ancient spatial information to reconstruct spatio-temporal landscape dynamics (Goodchild, 1991; Leyk et al., 2005). It allows quantifying the inherited part of SVE in the landscape variables derived from each planimetric data individually. It also offers the opportunity of investigating its impact on the calculation of dynamic landscape variables derived from several sources, which further limit error propagation in computing models. Beyond the scope of spatial modeling, the originality of PickShift lies in its ability to consider each source of uncertainty together: the production-, transformation-, and application-oriented uncertainty (Goodchild, 1991; Leyk et al., 2005). The two first sources are simultaneously considered in the bias calculated from GCPs (Goodchild, 1991; Leyk et al., 2005). Quantifying the respective contribution of coregistration and inherent quality source in the positional error estimates is a challenging task. The third source is considered through the calculation of indicators of

relative uncertainty (Goodchild, 1991; Leyk et al., 2005). They can be used in the decision process by users to measure the conformity of derived estimates regarding initial purposes. Furthermore, we suggest that future developments of PickShift should focus on the incorporation of different, spatially variable error components, such as digitization. Connections with existing tools to extract information from ancient spatial sources should also be considered (Plugin QGIS Historical Map, <https://plugins.qgis.org/plugins/HistoricalMap/>; Package R HistMapR, <https://github.com/AGAuffret/HistMapR>; Package Python HMVT, <https://github.com/EVS-GIS/hmvt-cli>; Herrault et al., 2013; Auffret et al., 2017; Dunesme et al., 2022, respectively). These PickShift enhancements would pave the way for a global analysis of geometric uncertainties in surficial estimates derived from ancient spatial sources.

We argue that PickShift could become a new paradigm in the processing of HPD. First, through its simplicity and availability, we expect more systematic measurements of spatial error and positive outcomes in historical land-cover studies. Second, we expect higher precision and transparency in the derived estimates. Third, we expect an increasing use of HPD since that can be accompanied by information about the quality and the reliability of output databases. Further bibliographical reviews and download analysis should be achieved to validate these expectations.

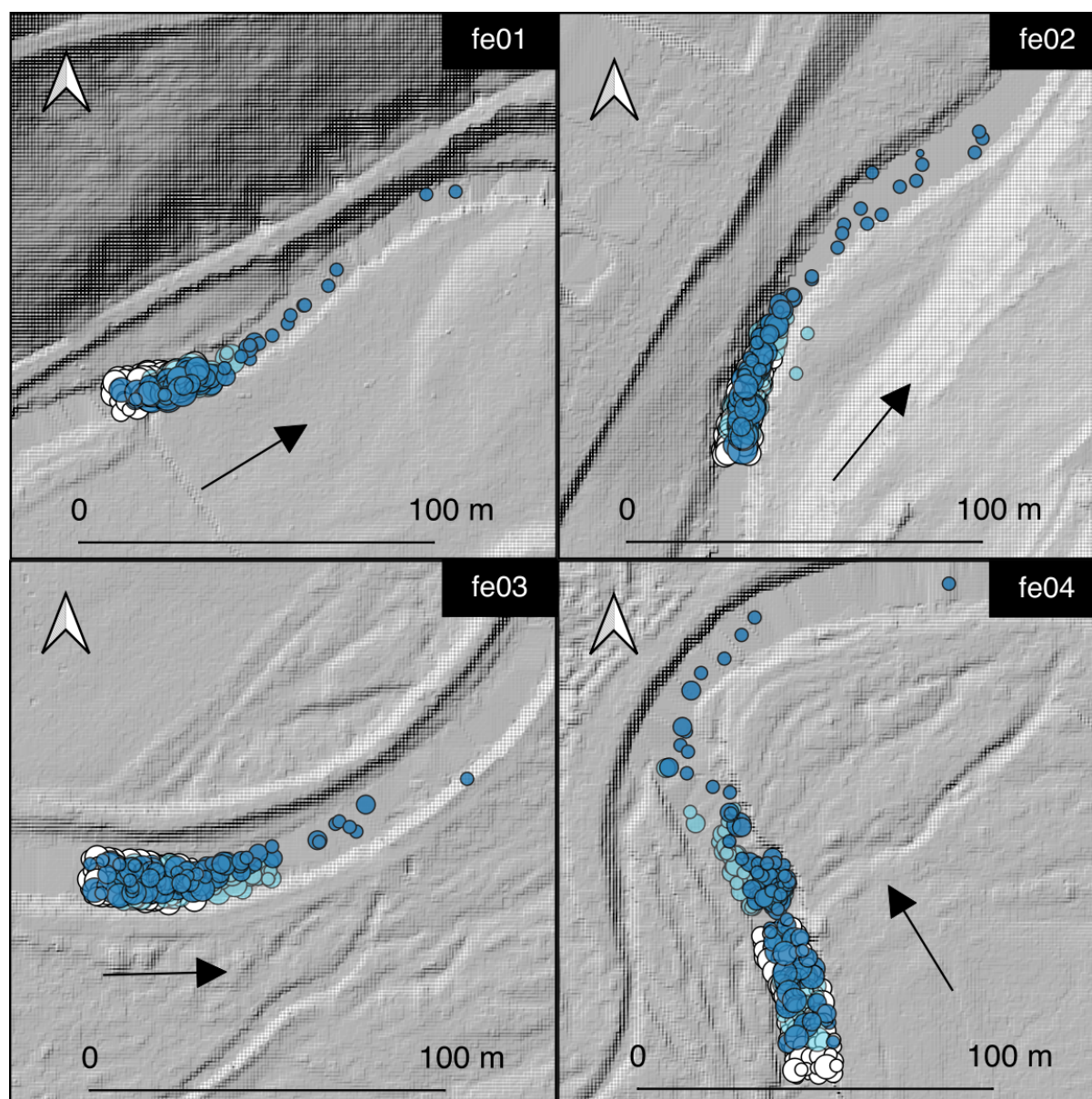
B.5 Conclusions

The use of historical planimetric data is essential for reconstructing landscape dynamics and anticipating future environmental changes. With the increasing availability of historical planimetric data and open-source GIS software's, academic expertise is no longer required to use these data. However, the spatially variable error that inherently affects these data must be carefully considered. If not, the measured surface changes may be misinterpreted. The user-friendly Python tool PickShift addresses this issue: it allows (1) computing and spatializing the geometric error associated with historical planimetric data, and (2) quantifying the surficial uncertainty associated with any digitized features. PickShift is also a versatile tool. On the one hand, its simple design allows using it without advanced GIS skills (e.g. environmental managers or students). On the other hand, its open code allows technical enhancements for more sophisticated studies encompassing the different facets of uncertainty in spatial models, such as (i) propagating uncertainty, (ii) modelling uncertainty on model inputs, (iii) resulting uncertainty on model outputs, and (iv) estimating variance-based sensitivity indices. We finally argue that PickShift can contribute to the generalization of methods for assessing uncertainties in the analysis of historical planimetric data and improve our understanding of spatio-temporal landscape dynamics, thereby contributing to more efficient environmental management practices.

Annexe C

Matériel supplémentaire pour le Chapitre 5

C.1 Cartographie des levés RFID



Granulométrie (axe-b, mm) :

Position initiale	Levé 1	Levé 2
◦ 16 - 32	◦ 16 - 32	◦ 16 - 32
○ 32 - 64	◦ 32 - 64	◦ 32 - 64
○ 64 - 128	◦ 64 - 128	◦ 64 - 128
○ 128 - 256	◦ 128 - 256	◦ 128 - 256
○ 256 - 1000	◦ 256 - 1000	◦ 256 - 1000

Sens de l'écoulement



FIGURE C.1: Cartographie de la position des particules équipées de traceurs RFID sur les sites de la Fecht, lors de leur mise en place et à chaque levé. Les sites sont classés d'amont en aval.

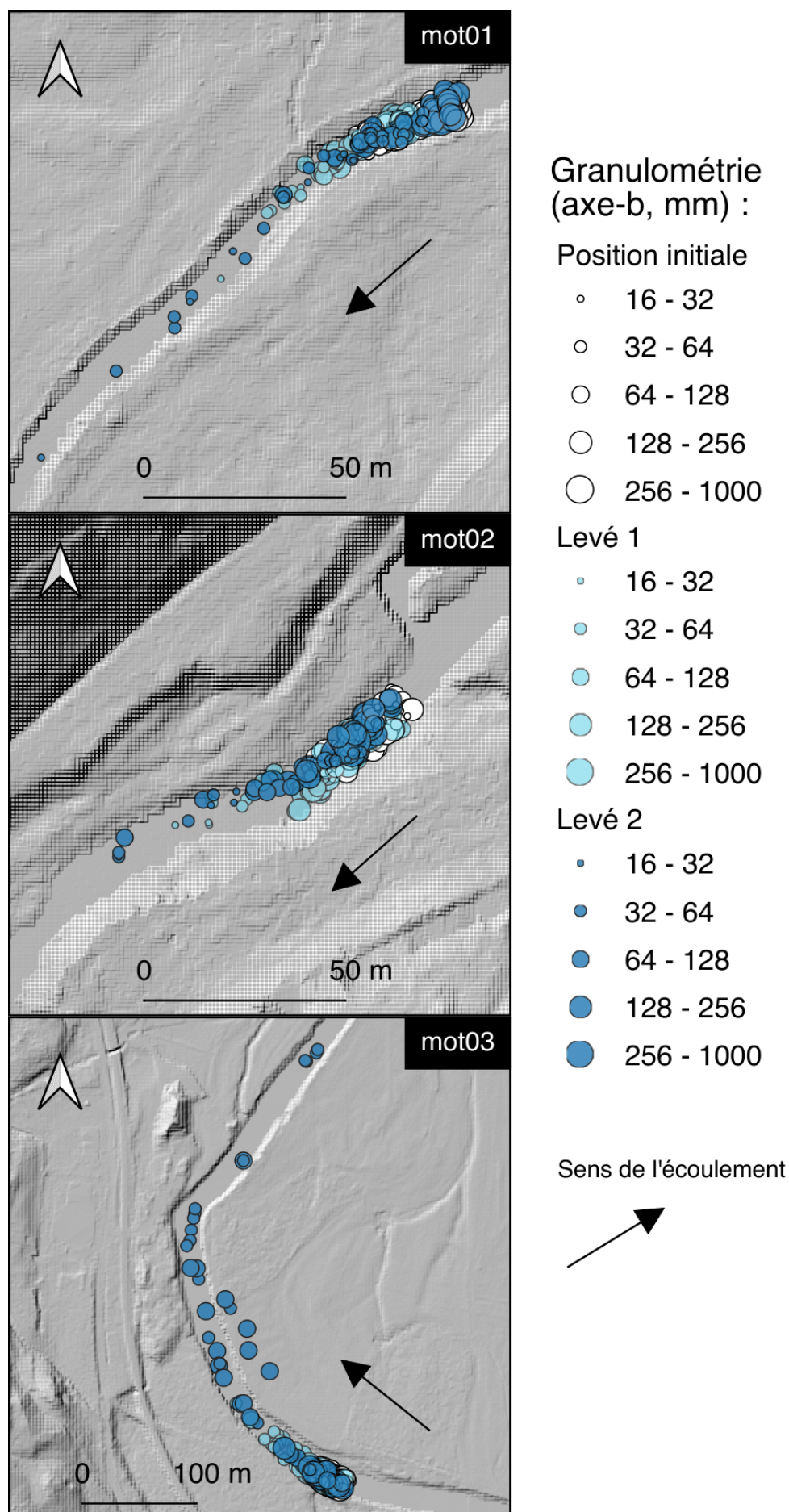


FIGURE C.2: Cartographie de la position des particules équipées de traceurs RFID sur les sites de la Moselotte, lors de leur mise en place et à chaque levé. Les sites sont classés d'amont en aval.

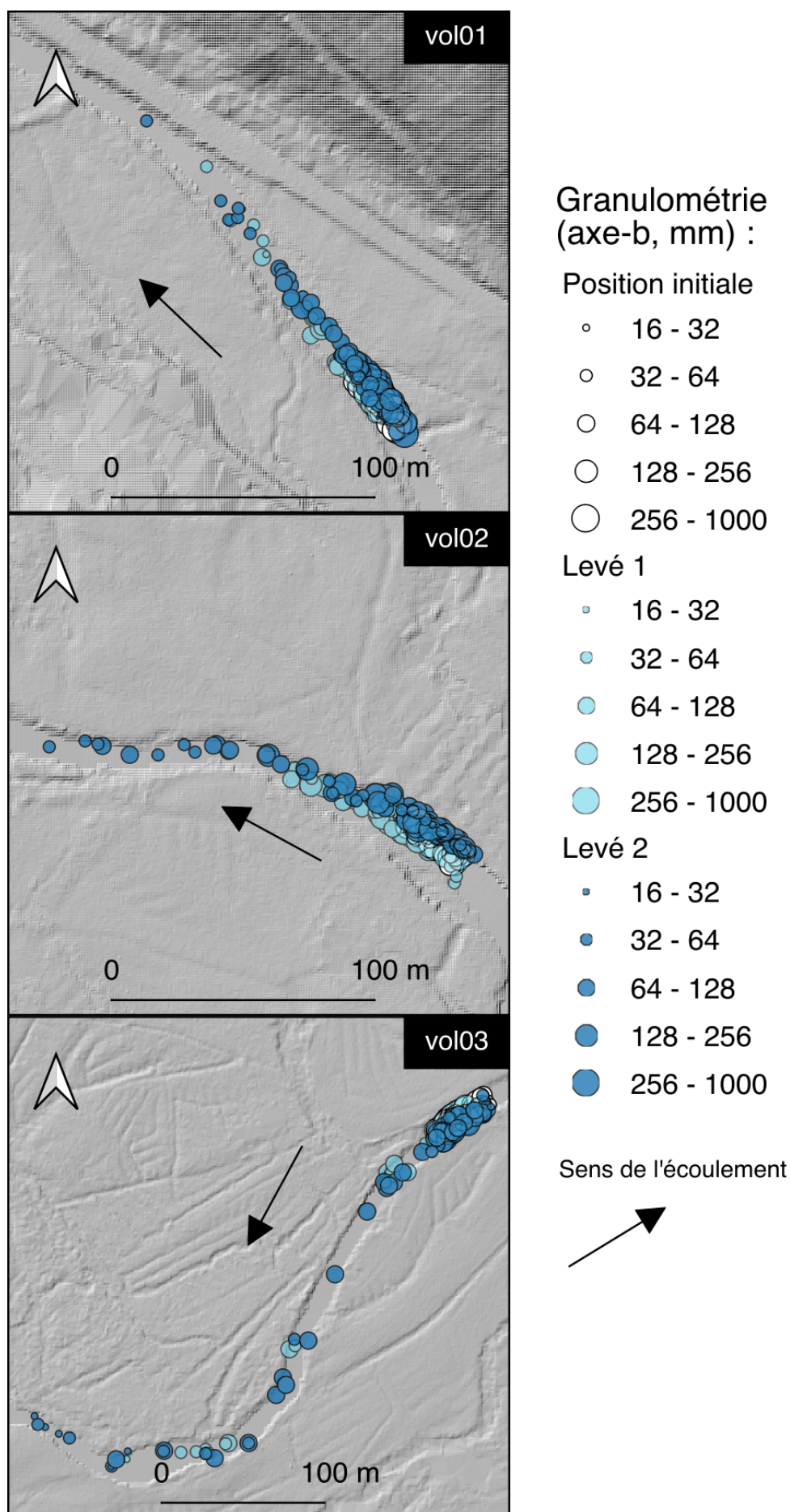


FIGURE C.3: Cartographie de la position des particules équipées de traceurs RFID sur les sites de la Vologne, lors de leur mise en place et à chaque levé. Les sites sont classés d'amont en aval.

C.2 Caractérisation des conditions de stabilité de tronçons à potentiel de réintroduction de la Mulette perlière (*Margaritifera margaritifera*) dans le bassin de la Vologne

Le bassin versant de la Vologne est connu pour avoir abrité jusqu'au début du XX^{ème} siècle d'importantes populations de Mulettes perlières (*Margaritifera margaritifera*). Actuellement, cette espèce est considérée comme écologiquement éteinte dans le bassin versant de la Vologne. En plus de la surpêche pour l'exploitation de ses perles (Fig. C.4B), les principales causes de son extinction sont attribuées aux problèmes de qualité de l'eau et à la dégradation du fonctionnement hydromorphologique de la Vologne et de ses affluents, qui résulte directement de l'industrialisation de la vallée. En réponse à cette situation, la Société d'Histoire Naturelle et d'Ethnographie de Colmar (SHNEC) a élaboré en 2021 un Programme de Conservation et d'Actions (PCA, Umbrecht, 2020), avec le soutien financier de l'Agence de l'eau Rhin-Meuse (AERM), de la Communauté d'Agglomération de Saint-Dié-des-Vosges (CASDDV), de la Communauté de Communes de Bruyères-Vallons des Vosges (CCB2V) et de la Direction Régionale de l'Environnement, de l'Aménagement et du Logement Grand Est (DREAL Grand Est).

Les trois objectifs du PCA sont les suivants (Umbrecht, 2021) :

- Atteindre un très bon état de fonctionnement écologique, au sens de la DCE pour l'ensemble des cours d'eau du bassin versant de la Vologne;
- Rétablir la fonctionnalité globale des écosystèmes fluviaux et riverains avec la conservation et/ou le rétablissement du cortège d'espèces associées *sensu lato*;
- Rétablir une population fonctionnelle de Mulette perlière.

La Mulette perlière est une espèce parapluie (Umbrecht, 2021). Elle exploite une niche écologique très étroite, en grande partie conditionnée par le substrat du lit des cours d'eau dans lequel l'individu juvénile vient s'insérer, qui requiert notamment une certaine stabilité (Hastie et al., 2000). Afin de caractériser les conditions de stabilité de la charge de fond des sites sélectionnés par la SHNEC pour une potentielle réintroduction de la Mulette perlière, une collaboration a été entreprise avec ce travail de thèse. Trois sites ont ainsi été équipés de traceurs RFID, dans deux affluents de la Vologne : le Barba et le Neuné (Fig. 5.7). Ces deux affluents étaient historiquement largement peuplés par la Mulette perlière.

L'hypothèse de la SHNEC est que le Barba reste en l'état -sur des critères uniquement physico-chimique- l'affluent le plus favorable du bassin versant pour une tentative de réintroduction. Il convenait donc de tester également la stabilité du substrat (e.g. Vaessen et al., 2021). Par

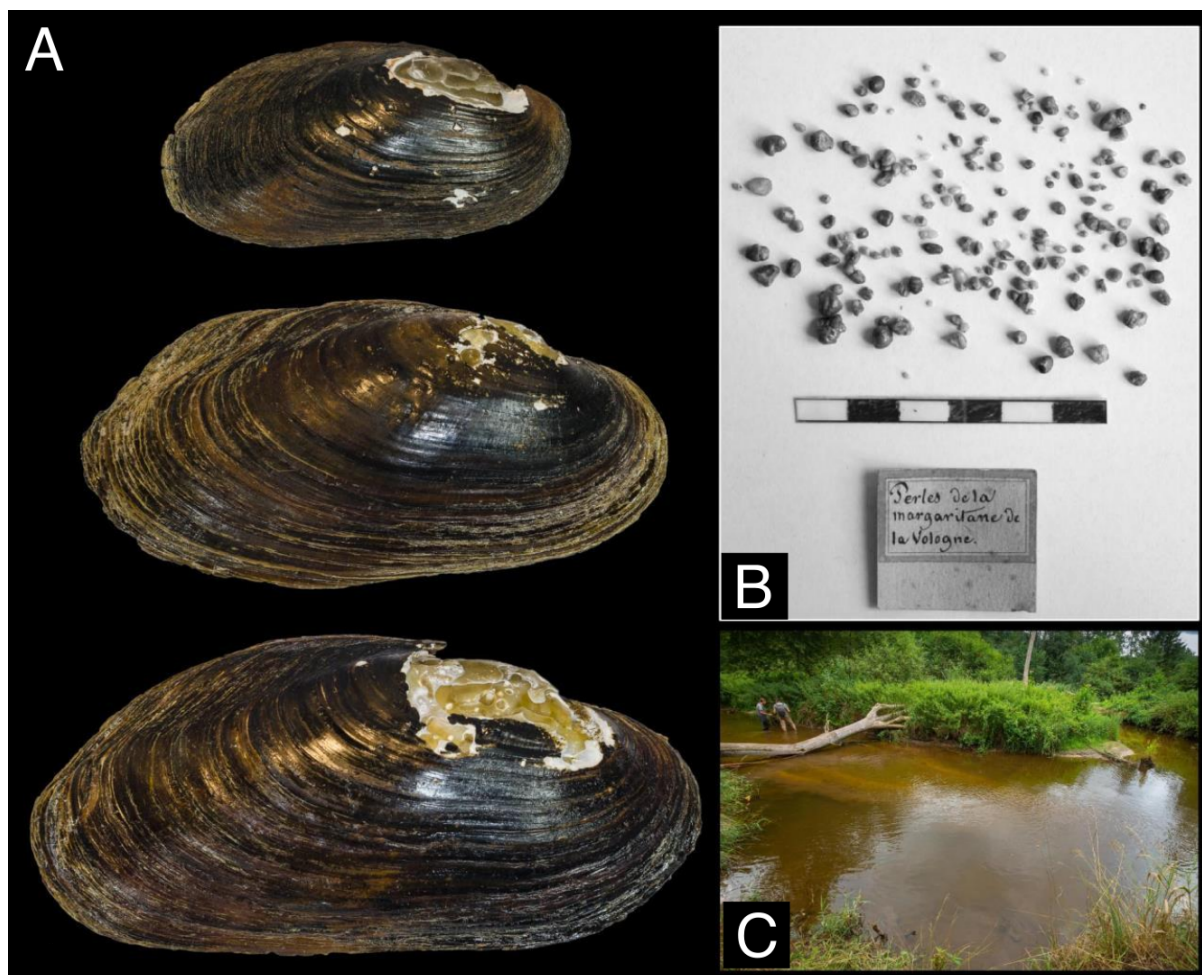


FIGURE C.4: Mulette perlière dans le bassin versant de la Vologne. (A) Spécimens récoltés par le Dr. Mougeot dans la Vologne et conservés au Musée d'Histoire naturelle et d'Ethnographie de Colmar (photos : J.-M. Bichain). (B) Perles prélevées sur des Mulettes perlières de la Vologne, collection du Muséum-Aquarium de Nancy (photo : M. Chabrol). (C) Habitat à Mulette perlière sur le Neuné jusque dans la deuxième moitié du XX^{ème} siècle (photo : J. Kaufmann). Tiré de Umbrecht (2020).

conséquent, l'objectif de cette collaboration est de comparer la stabilité de la charge de fond du Barba et du Neuné, afin de vérifier l'hypothèse de la SHNEC. Un total de 200 traceurs RFID ont été mis en place dans les alluvions du Barba (2x50) et du Neuné (100). Deux levés ont été réalisés (Fig. C.5). Les résultats préliminaires indiquent une différence significative de transport de la charge de fond entre les deux affluents. Entre le 1^{er} et le 2^{ème} levé, un total de 14% et 30% d'alluvions ont été transporté sur le Barba et le Neuné, respectivement (Fig. C.6). En plus d'une granulométrie plus grossière sur le Barba que sur le Neuné, les distances de transport des quelques alluvions mobilisés sur le Barba sont également plus faibles (<25 m) que sur le Neuné.

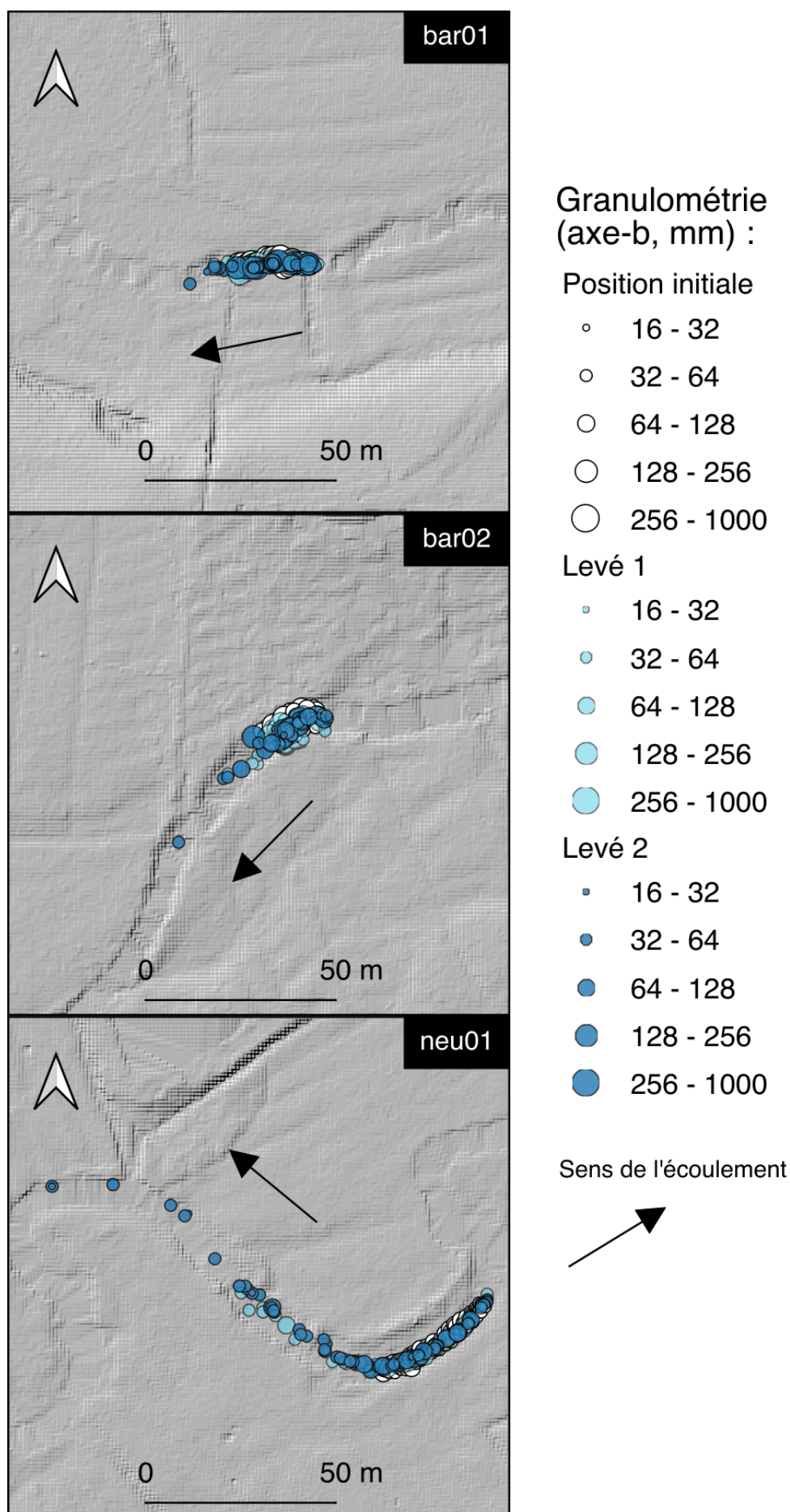


FIGURE C.5: Cartographie de la position des particules équipées de traceurs RFID sur deux affluents de la Vologne (Barba et Neuné), lors de leur mise en place et à chaque levé.

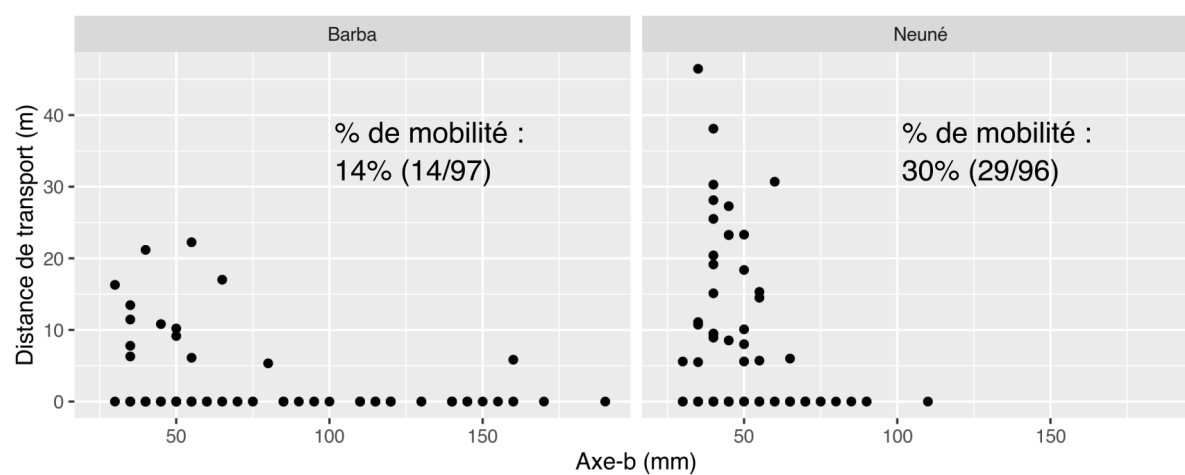


FIGURE C.6: Distance de transport des particules en fonction de leur axe-b, entre le 1^{er} et le 2^{ème} levé. La proportion de traceurs 'mobiles' a été défini sur la base d'une distance de transport minimale de 5 m.

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Résumé

La dynamique morpho-sédimentaire du massif des Vosges reste mal connue, tant concernant les flux sédimentaires actuels que ceux des stocks sédimentaires anciens. Or, le massif des Vosges présente des caractéristiques uniques : forts contrastes lithologiques, englacement passé partiel, et densité de population élevée. La thèse évalue ainsi la dynamique morpho-sédimentaire des rivières vosgiennes à travers trois objectifs : mesure des taux de dénudation pluri-millénaires via des analyses cosmogéniques (^{10}Be et ^{26}Al), étude des ajustements fluviaux post-glaciaires d'un bassin versant anciennement fortement englacé, et étude de l'impact de la pression anthropique sur les ajustements fluviaux au cours des deux derniers siècles. Les résultats démontrent une influence significative des héritages glaciaires et anthropiques sur la dynamique actuelle des rivières, et soulignent l'importance de prendre en compte ces facteurs pour élaborer des stratégies de gestion et de restauration efficaces.

Mots clés : Vosges, dynamique morpho-sédimentaire, dénudation, héritages glaciaires, pression anthropique

Résumé en anglais

Morpho-sedimentary dynamics in the Vosges Massif remain poorly understood, both in terms of current sedimentary flows and ancient sedimentary archives. Yet the Vosges massif offers unique characteristics: strong lithological contrasts, partial glaciation and high population density. The thesis therefore assesses the morpho-sedimentary dynamics of Vosges rivers through three objectives: inferring multi-millennial denudation rates via cosmogenic analyses (^{10}Be and ^{26}Al), studying post-glacial fluvial adjustments in a formerly heavily glaciated catchment, and investigating the impact of anthropogenic pressure on fluvial adjustments over the last two centuries. The results demonstrate a significant influence of glacial and anthropogenic legacies on current river dynamics, and underline the importance of taking these factors into account to develop effective management and restoration strategies.

Keywords : Vosges, morpho-sedimentary dynamics, denudation, glacial legacies, anthropic pressure