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Geomorphology of the Hells Canyon Reach of the Snake River

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Hells Canyon Complex
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DEFINITIONS

A	Cross-sectional area
Alkalic	A group of igneous rocks that contain relatively high proportions of sodium and potassium
Aphanitic	A characteristic of fine-grained igneous rocks in which crystals are not visible to the naked eye
Arcuate	Curved or bowed in shape
Argillite	A compact metamorphic rock derived from mudstone or shale that is relatively indurate
Barbed	A drainage pattern produced by tributaries that join the mainstem in sharp bend that point upstream; this pattern is usually the result of stream piracy that has reversed the direction of the mainstem flow
Base level	The theoretical limit toward which erosion of the earth surface constantly progresses, including the lowest level below which a stream cannot erode its bed
CRBs	Columbia River Basalts
Clastic	A sedimentary rock that is principally derived from fragments of pre-existing rocks that have been transported from their place of origin
Country rock	The pre-existing rock that is intruded or traversed by an igneous intrusion or mineral deposit
D	Flow depth
d_{50}	Median grain size
d_{50s}	Median surface grain size
d_{50ss}	Median subsurface grain size
d_n	Grain size for which *n* percent of the sizes are smaller
D_s	Scour depth
Dikes	An igneous body that cuts across the structure of adjacent rocks or cuts massive rocks

Diorite	A group of plutonic rocks that are intermediate in composition between acidic and basic
Disjunct species	occurrence of a plant population that is geographically removed from the main distribution of the species
Dry ravel	The downhill movement of soil and debris during dry periods caused by gravitational forces
Duripan	A soil horizon characterized by silica cementation (e.g., hardpan)
Foliated	A planar, banded arrangement of textural or structural features in any type of rock
G	Gravitational constant
Graben	A depressed block that is bounded by faults
Grandioritic	Characterizing a coarse-grained plutonic rock that is intermediate in composition between a quartz diorite and quartz monzonite
Grus	An accumulation of angular, coarse-grained fragments of crystalline rocks (especially granite) in an arid or semi-arid region
h^*	Relative submergence (R/d_{50})
Headcutting	The lengthening of a valley by erosion at the valley head
Isocline	A fold whose limbs are parallel
Indurated	Characterisitc of a rock or soil that has been hardened or consolidated by pressure, cementation, or heat
Intermontaine	Lying between mountains
κ	von Karman's constant
K-spar	A class of feldspars minerals that are potassium-rich and can be identifiable by their pinkish hue
Lacustrine	Pertaining to, produced by, or inhabiting a lake environment
Loess	A blanket deposits of buff-colored wind-blown silt that is homogenous and non-stratified
Mica	A group of complex silicates with perfect basal clearage that split into thin elastic laminae sheets and vary from colorless to black

N	Roughness coefficient
Olivine	A green or brown mineral common to low-silica igneous rocks (e.g., basalt)
Orographic	Related to the processes and resultant features of mountain-building
ρ	Fluid density
ρ_s	Sediment density
Paleoclimate	A climate from the geologic past
Peneplane	A low, nearly featureless land surface that has been produced by long-term mass wasting, sheetwash, and stream erosion processes
Phenocryst	A relatively large and conspicuous crystal that is visible within an igneous rock
Plagioclase	A relatively common class of feldspars minerals that ranges from albite (sodium-rich) to anorthite (calcium-rich)
Pluton	An igneous intrusion
Q	Hydraulic discharge
$Q_{1.5}$	Hydraulic discharge with a 1.5-year recurrence interval
Q_{100}	Hydraulic discharge with a 100-year recurrence interval
q^*	Dimensionless hydraulic discharge per unit width
Qs	Sediment discharge
Q_b	Bedload sediment discharge
$q_b{}^*$	Dimensionless bedload transport rate
$q_b{}^*{}_D$	Dimensionless bedload transport index
R	Hydraulic radius
R_{sf}	Hydraulic radius pertaining to the skin friction of the bed material
$\boldsymbol{R}$	Submerged specific gravity of sediment $[(\rho_s-\rho)/\rho]$
Residual pool depth	Difference between pool bottom and downstream riffle crest

S	Energy slope
S	Channel slope
S_s	Volume of sediment stored
Self-formed	Morphological features that are not forced by external controls
Shield volcano	A broad, gently sloping volcanic cone that is usually several tens or hundreds of square miles in extent
Solifluction	The slow, downslope movement of water logged soil that is particularly common at high elevations in regions underlain by frozen ground that acts as a downward barrier to water percolation
Suture	A boundary line, or line of contact
τ_0	Total boundary shear stress determined as a depth-slope product
τ^*_{50s}	Dimensionless shear stress of the median surface grain size (d_{50s})
τ^*_{50ss}	Dimensionless shear stress of the median subsurface grain size (d_{50ss})
τc_{50s}	Critical shear stress of the median surface grain size (d_{50s})
τc_{50ss}	Critical shear stress of the median subsurface grain size (d_{50ss})
$\tau^* c_{50s}$	Dimensionless critical shear stress for incipient motion of the median surface grain size (d_{50s})
$\tau^* c_{50ss}$	Dimensionless critical shear stress for incipient motion of the median subsurface grain size (d_{50ss})
τ_{sf}	Skin friction stress (that portion of the total boundary shear stress acting on the bed and responsible for sediment transport)
Thrust fault	A fault with a dip of 45 degrees or less over much of its extent, on which the hanging wall appears to have moved upward relative to the footwall. Horizontal compression rather than vertical displacement is its characteristic feature.
u	Average velocity
$\bar{u}$	Local average velocity
$\langle u \rangle$	Vertically-averaged velocity determined from the law of the wall
u^*	Shear velocity

V	Flow velocity
Volcaniclastic	A sedimentary clastic rock containing volcanic material
W	Top flow width
$W_{1.5}$	Flow width associated with the 1.5-year hydraulic discharge ($Q_{1.5}$)
W_{100}	Flow width associated with the 100-year hydraulic discharge (Q_{100})
z_0	Height above the bed where the velocity profile goes to zero

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ACRONYMS

BLM	Bureau of Land Management
BMPs	Best Management Practices
BPA	Bonneville Power Administration
CRB	Columbia River Basin
DFNWR	Deer Flat National Wildlife Refuge
EPA	U.S. Environmental Protection Agency
FERC	Federal Energy Regulatory Commission
GLO	Government Land Office
HCC	Hells Canyon Hydroelectric Complex
HCNRA	Hells Canyon Natural Recreation Area
IDAPA	Idaho Administrative Procedures Act
IDEQ	Idaho Department of Environmental Quality
IDWR	Idaho Department of Water Resources
IPC	Idaho Power Company
NCASI	National Council for Air and Stream Improvement
NOAA	National Oceanic and Atmospheric Administration
NPPC	Northwest Power Planning Council
POR	periods of record
SCS	Soil Conservation Service
USACE	United States Army Corps of Engineers
USDA	United States Department of Agriculture
USFS	United States Forest Service
USGS	United States Geological Survey

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EXECUTIVE SUMMARY

The purpose of this report is to assess how the construction and operation of Idaho Power Company's Hells Canyon Hydroelectric Complex (HCC) affects the local downstream geomorphology in Hells Canyon. In a current physiographic context, the Snake River flows through the Snake River Plain and immediately upstream from the HCC, five major tributaries (the Owyhee, Boise, Malheur, Payette, and Weiser rivers) join the Snake River as it flows northward along the Idaho–Oregon border. Below the HCC, Hells Canyon is characterized by a deep, narrow gorge with essentially no active floodplain that was created by geologic structural development and rapid downcutting. Numerous local tributaries within the canyon drain a variety of basin areas, and the mainstem Snake River is joined by the Imnaha and Salmon rivers near the downstream boundary of the study area.

The geologic history of the canyon is complex and the Snake River began significantly cutting Hells Canyon around 2.0 to 2.5 million years ago. During the Quaternary glacial period the canyon continued to be shaped by sustained seasonal flows estimated to have been at least 10 times larger than discharges of present streams. In addition, about 14,500 years ago a catastrophic flood from Lake Bonneville escaped into the Snake River Plain and Hells Canyon. The resulting pre-historic hydrology and floods formed the bed and channel form of the Snake River above and below the HCC that is seen today. During the last 10,000 years, hydraulic discharges and sediment transport capacity have markedly decreased toward current conditions. Regional climate conditions have stabilized within the last approximately 1,000 years to a state similar to current conditions, which has not produced floods of the magnitude of those experienced previously. As a result, the Snake River channel in Hells Canyon has experienced little change since it was formed, and is very stable under recent flows.

Since the geologic framework for the Snake River system was put into place, more recent anthropogenic disturbances have impacted physical process in the study area in general, and in Hells Canyon specifically. By the 1880s land uses varied by river reach, but overall these activities likely substantially increased sediment supplies relative to pre-settlement conditions. These land uses primarily included trapping, mining, forest management, wildfire, and agricultural development. Following most of these anthropogenic effects, water storage and regulation became the most significant anthropogenic disturbance in the basin with the greatest influence on the hydrology and sediment supply. Independent from the HCC, over 10 mainstem and 35 tributary facilities were constructed in the Snake River system between 1901 and 1969. By the time Brownlee Reservoir was completed in 1958, 87% of the upstream drainage area had already been cut off by other reservoirs, effectively keeping most of the upstream supply of sediment from entering Hells Canyon.

The timing of these disturbances is an important issue. In general, many of these activities, such as trapping, mining, timber production, livestock grazing, and agriculture, likely caused a general increase in sediment supply above pre-settlement conditions starting in the early 1800s. As such, although some of these activities, including grazing and irrigation, continue today, the historical sediment loads produced by these anthropogenic activities began to be trapped behind the numerous reservoirs in the basin beginning in the early 1900s. Thus, in the last 150 years,

anthropogenic disturbances throughout the watershed first caused significant additional sediment supply to the system and then subsequent decreases in sediment supply, in part, because of multiple water resource projects. The combination of these factors likely produced a “slug” of sediment that has either worked, or continues to work, its way through the system.

A number of important physical conditions affect the current geomorphology of the Snake River Basin and Hells Canyon. These influences include topography, climate, geology, hydrology, soils, and vegetation. The Snake River hydrograph is driven by a snow-melt regime with dynamic, flashy discharges in the tributaries. Rain-on-snow events are typically responsible for floods and debris flows, with spring runoff events accounting for the majority of annual sediment yields.

Stream flows and sediment dynamics are the primary influences on geomorphology in the Snake River system. Water and sediment that enters into Brownlee Reservoir is directly related to the processes at work in the mainstem reach immediately above the HCC, which in turn is influenced by inputs from the next upstream mainstem reach and the Owyhee, Boise, Malheur, Payette, and Weiser rivers. Numerous other upstream regulation projects have altered stream flows; during dry years, almost half of the estimated naturally-occurring volume of the Snake River is diverted for agricultural purposes. Most importantly from a geomorphic context, these projects have also reduced peak flows rates that transport the majority of sediment load. Thus, the available sediment transport capacity has continued to decline compared to historical (geologic) conditions.

Upstream from the HCC, the Snake River is located in a relatively flat alluvial plain, where coarse sediment that reaches the mainstem is generally deposited prior to reaching the HCC. A combination of upstream regulation projects and a flat gradient have caused channel islands upstream from the HCC to increased in areal extent by an average of 8% since 1938. Sediment that does reach the HCC from the upstream reach is comprised almost entirely of silt-clay and very fine sand sediments. In the absence of the HCC, these fine-grained materials likely would flush through the Hells Canyon reach. Coarser sand and gravels in the headwaters of Brownlee Reservoir and from local tributaries appear to be only a minor portion of the total downstream coarse sediment supply and only a small fraction of pre-regulation sediment supplies in the basin.

Downstream from the HCC, the channel in Hells Canyon is controlled, in large part, by the imprint of its complex geologic history. Most hillslope, valley, and channel morphology features appear to be relic features associated with pre-regulation hydrologic and geologic events (i.e., prior to the HCC, as well as prior to the numerous other regulation projects in the upper Snake River Basin). Although pre-settlement and pre-regulation conditions are as not well known as current conditions, the weight of evidence suggests that Hells Canyon has been a largely static river system for at least 1,000 years, if not longer.

Current peak discharges are not markedly different from pre-HCC flows because the HCC is capable of storing only approximately 11% of the river’s average annual flow. Therefore, the HCC has relatively little ability to regulate floods because the reservoirs fill so rapidly. Furthermore, pre- and post-regulation releases are not of sufficient magnitude to mobilize the erosion-resistant geomorphic features in Hells Canyon that exert significant hydraulic control on the river.

In Hells Canyon, the river morphology is characterized by a deep, steep, and narrow valley that is confined by bedrock walls, talus slopes, debris flows, landslides, and alluvial terraces (Bonneville Flood deposits, landslide backwater deposits, relict alluvial fans, and relict bars). This confinement has precluded the development of an active alluvial floodplain that is typical of other rivers of comparable discharge area and gradient. Much of the river morphology is forced by large-scale geologic and geomorphic controls that significantly reduce the range of fluvial processes and types of channel adjustment found in other alluvial rivers. For example, rapids and pools caused either by boulder bed material or by debris-flow fans are common in the canyon.

Local tributaries and adjacent hillslopes together are estimated to produce a minimum of 8.6 million tons of sediment on an annual basis. While the small tributaries within Hells Canyon are small in comparison to the larger upstream tributaries, their influence on local processes of sediment production, transport, and deposition is significant because they are linked directly with the mainstem Snake River in this reach. These sources are also significant relative to the reduced supply from the larger upstream tributaries that have been cut off by other regulation projects. Sediments are supplied directly from the tributaries on relatively short timescales (10s to 100s of years) during peak flow events. Additional sediment accumulates within most of these tributaries above sharp bends upstream from their confluence with the Snake River. These materials likely are mobilized only during extreme events on longer timescales (100s to 1000s of years). Steep hillslopes (62% have a slope greater than 40 degrees) have also produced sediment directly to the mainstem over the last 1,000 years, as well as over longer geologic timescales. Rock varnish confirms that episodic small to catastrophic massive landslides continue to occur along both sides of Hells Canyon.

Riverbed material does not appear to produce substantial volumes of sediment and is not extensively reworked during peak events because the channel is largely a stable, armored system. Based on weathering patterns on the armored gravel bars, these features appear to have been armored for a significant period of time (on the order of 100s to 1000s of years), and not as a result of the HCC.

Once sediments reach the mainstem from the tributaries and hillslopes, the issue of relative sediment transport capacity and sediment supply becomes critical. The very large transport capacity of the mainstem appears to have been so effective that the enormous volumes of sediment produced by local sources have not been sufficient to preclude rapid downcutting, as is evidenced by the very existence of the young, narrow, steep canyon. Thus, the mainstem channel has apparently been generally sediment deficient with respect to the upstream and local supply of sediment over a long geologic timescale, independent from the construction of the HCC.

In Hells Canyon large-scale geologic and geomorphic processes control the overall channel morphology and these processes are not appreciably altered by the HCC. Many of the potential channel responses to reduced sediment inputs simply cannot compete with the larger geologic controls at work in the canyon. For example, decreased sediment loads generally cause rivers to armor their beds and decrease the zone of active sediment transport. In this case, the Snake River channel bed in Hells Canyon likely was already well-armored before the HCC was built, so the expected response is likely to be muted. Although isolated pockets of materials (sand bars, terrace banks) are subject to respond to the minor decrease in fine-grained upstream sediment load, it is likely that these features are fluctuating within a much larger dynamic associated with the complex geologic history of the canyon.

Recent XRD and grain size data suggest that sandbars in Hells Canyon were formed primarily from upstream materials that were deposited within the canyon more than 2,500 years ago. Since the original depositional event(s), the watershed has not experienced flood conditions that changed the original depositional signature. XRD analysis suggests that the Idaho Batholith was an important source of material for the original sandbars formed in Hells Canyon. This source was cut off by water storage projects prior to and upstream from the HCC. While the HCC has cut off supplies of sediment from Hells Canyon, grain-size data suggest that Brownlee Reservoir has not trapped sediments in the size range necessary to maintain sandbars in the condition observed prior to construction of the HCC. It is unclear what effect the HCC has had on accelerating the erosion of sandbars downstream from HCD within long-term and short-term dynamic cycles of erosion and degradation associated particularly with canyon environments.

PREFACE

The Hells Canyon Complex is located within the Snake River Basin, which has been the focus of numerous studies and reports over the last 30 years. As part of this process, the Interior Columbia Basin Ecosystem Management Project (ICBEMP) has compiled a large database of land attributes that is available to the public. **With the exception of the local geology in Hells Canyon, all of the data used to generate basin-wide summaries were supplied by the ICBEMP dataset. This includes all of the Geographic Information System (GIS) figures and all of the summary tables that present such data as land use, grazing allotments, vegetation, and similar information.** Local geology in Hells Canyon was mapped using an IPC GIS dataset that was based on more detailed geologic mapping conducted by the University of Idaho and Tracy Vallier with the U.S. Geological Survey.

Photographs from Hells Canyon were either aerial oblique images from a September 2000 fly-over or color aerial photos taken at a scale of 1:8400 on August 9, 1997.

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1. INTRODUCTION AND GEOLOGIC AND GEOMORPHIC HISTORY

Chapter Summary

The purpose of this report is to assess how the construction and operation of Idaho Power Company's (IPC's) Hells Canyon Hydroelectric Complex (HCC) affects the local downstream geomorphology (Section 1.1.). The first page of each chapter begins with a summary, as this one does, to provide an overview of the material and key conclusions. Then, the chapter subsections contain a detailed description of the data supporting those conclusions.

Chapter 1 contains a description of the geologic and geomorphic history of Hells Canyon, which contains the three hydropower facilities in the HCC. In a current physiographic context (Section 1.2.), the Snake River flows through the Snake River Plain and immediately upstream of the HCC, five major tributaries (the Owyhee, Boise, Malheur, Payette, and Weiser rivers) join the Snake River as it flows northward along the Idaho–Oregon border. Hells Canyon proper is characterized by a deep, narrow gorge with no true floodplain that was created by geologic structural development and rapid downcutting. Numerous local tributaries within the canyon drain a variety of basin areas, and the mainstem Snake is joined by the Imnaha and Salmon rivers near the downstream boundary of the study area.

As described in the pre-Quaternary geologic history (Section 1.3.), the walls of Hells Canyon are composed of hard, brittle, dense formations that were metamorphosed under extreme temperatures and pressures. Canyon rocks also include thick basalt formations collectively known as the Columbia River Basalts (CRBs). The Snake River began significantly cutting Hells Canyon around 2.0 to 2.5 million years ago by draining large volumes of water during glacial periods and by tectonic uplift along the many faults in the canyon. During the Quaternary glacial period (Section 1.4.) the canyon continued to be shaped by sustained seasonal flows estimated to have been probably at least 10 times larger than discharges of present streams. In addition, about 14,500 years ago a catastrophic flood from Lake Bonneville escaped into the Snake River Plain and Hells Canyon. During the last 10,000 years, hydraulic discharges and sediment transport capacity have markedly decreased toward current conditions.

Understanding the geologic history of the Snake River Plain and Hells Canyon is essential to interpreting current geological conditions and potential sediment delivery through the system (Section 1.5.). Many of the high-producing sediment areas are located either in the headwaters of major tributaries to the Snake River, or in the steep local tributaries within Hells Canyon itself. Structural geology, particularly the faults in Hells Canyon, have also contributed to sediment supplies; for example, the relatively straight course of the Snake River through Hells Canyon is believed to be controlled mainly by north- to northeast-trending structural faults.

1.1. Introduction

Idaho Power Company (IPC) prepared this assessment of how the construction and operation of IPC's Hells Canyon Hydroelectric Complex (HCC) affects the local geomorphology in Hells Canyon. The purpose of the assessment is to provide information to IPC to assist in relicensing these facilities with the Federal Energy Regulatory Commission (FERC).

The structure of this report reflects the study approach:

1. Describe the geologic and geomorphic history—Chapter 1
2. Describe the anthropogenic influences on the geomorphology of the study area—Chapter 2
3. Conduct an inventory of current conditions—Chapter 3
4. Describe the physical processes upstream of Hells Canyon Dam—Chapter 4
5. Describe the physical processes downstream of Hells Canyon Dam—Chapter 5
6. Summarize the contribution of the HCC to the current stream flows, geomorphology, and sediment dynamics in Hells Canyon—Chapter 6

Throughout this report, the significant points contained in each chapter are summarized at the beginning of each chapter. Then, chapter subsections contain a detailed description of the data supporting those conclusions. The focus of the study is to determine the effects of natural and anthropogenic factors on geomorphic conditions. Various historical (geologic) and current factors that influence geomorphology are described only to the extent needed to address their effects.

This chapter provides an overview of the geologic and geomorphic history of the Snake River Basin, including Hells Canyon. The study area in this report includes the Snake River from its headwaters near the Idaho–Wyoming–Montana border downstream to the confluence with the Clearwater River near Lewiston, Idaho, at river mile (RM) 139 (268,300 km^2 [103,600 mi^2]; Figure 1.1). Figure 1.2 provides a schematic of the study area, showing the locations of major tributaries and reservoirs.

The geology of the basin is described in detail because it directly affects the geomorphology through a series of distinct temporal and spatial scales (Figure 1.3). Basin-wide geological processes operating during more than 100,000 years affect available surficial materials, topography, and hydrology that control sediment production throughout the basin. Within this basin, the river valley is affected by geological conditions that reflect processes and events that have occurred over 1,000- to 100,000-year time scales. Finally, although the Snake River itself is influenced primarily by channel processes that operate over time frames of 10 to 1,000 years, these processes are somewhat limited by valley conditions that have been imprinted by the Bonneville Flood and Holocene flooding events.

The geologic history of the basin is important because the interaction between the river and the surrounding geology directly influences how the Snake River conveys water and sediment through its system. A physiographic description of the Snake River Basin is presented to provide the background for the study area of this report. Following this description, the pre-Quaternary geologic history of the Snake River Basin is summarized chronologically and includes the following major events and influences:

- Tectonic Formation of Hells Canyon Area
- Emplacement of Idaho Batholith
- Development of Basin and Range Province
- Eruption of the Columbia River Basalts (CRBs)
- Development of Snake River Plain
- Formation of Lake Idaho
- Capture of the Snake River and Incision of Hells Canyon

The Quaternary influences on present-day geomorphological features, including climate and flooding conditions and the Bonneville Flood, are also presented in this chapter. Anthropogenic activities have also affected the geomorphology of the basin and Hells Canyon. These disturbances, which include water regulation and storage, fur trapping, mining, forest management, grazing, agricultural development, fire, urbanization are discussed in Chapter 2.

Natural and anthropogenic events have had a large influence on the current configuration of the Snake River mainstem and its tributaries. Although much literature is available describing each of these geologic events and periods in detail, this overview focuses on aspects of these events that relate directly to an understanding of 1) current conditions of the Snake River Basin and Hells Canyon, and 2) the hydrology and sediment processes of the Snake River in and immediately above the Hells Canyon reach. Throughout this report, definitions are provided for unusual geologic terms or other vocabulary that denotes a specific technical meaning; these definitions are included in the front section of this report following the Table of Contents.

1.2. Current Physiographic Description

Within the study area, the Snake River and its tributaries drain approximately 268,300 km^2 (103,600 mi^2) in western Wyoming, southern and central Idaho, northwestern Utah, northern Nevada, eastern Oregon, and southeast Washington (Figure 1.1). The Snake River Basin encompasses most of Idaho south of Lewiston with the surface area of its drainage basin covering about 87% of the state. The basin includes the highest point in Idaho (Borah Peak, elevation 3,859 m [12,662 ft]) and the lowest point in Idaho (Lewiston, elevation 224 m [735 ft]).

Within the study area, the Snake River flows through a variety of physiographic features that form a continuum from its headwaters through the Snake River Plain and Hells Canyon. A generalized regional physiographic map is provided for reference in Figure 1.4. The Snake River Plain is a vast volcanic basin that is the dominant feature of southern Idaho. Hackett and Morgan

(1988) have characterized the plain as a "40-to 60-mile wide, arcuate belt that extends about 644 km (400 miles) from Yellowstone National Park to the Idaho–Oregon border." Based on numerous studies, the plain is generally further subdivided into an eastern and western portion due to differences in structure. Roughly east of Twin Falls, the eastern Snake River Plain is generally described as an active volcanic area with a flat expanse of lava flows from low shield volcanoes and fissures, dotted with occasional low cinder cones (Hackett and Morgan 1988). In places, a thin veneer of basalt caps thick deposits of rhyolite to form resistant buttes. In contrast, the western Snake River Plain has been characterized as a northwest-trending, lowland graben bounded by high terraces and numerous normal faults (Leeman 1982). Voluminous sedimentary deposits overlie thick basaltic and rhyolitic volcanic rocks in this area (Othberg 1994, IDWR 1981).

The Boise, Smokey, and Sawtooth mountain ranges in the central Idaho mountains form the northern boundary of the Snake River Plain. The mountain ranges are interspersed with large, intermontaine valleys such as the Little Lost River Valley. Historic widespread basaltic volcanism to the south of these ranges caused the Snake River to be displaced toward the southern margin of the plain (O'Connor 1993). Well-defined topographic features are lacking along the southern portion of the plain where smaller tributaries dissect the plain as it rises to merge with the mountains to the south (Mabey 1982). Along the southern margin in the western Snake River Plain, the Snake River is incised as much as 182 m (600 ft) below the surrounding basaltic uplands. Along the mainstem there are wide valleys with low gradients, formed where the river was displaced into erodable sediments. These valleys are generally interspersed with areas that are characterized by high gradient, narrow canyons where the river was forced to erode resistant volcanic rocks (O'Connor 1993). For example, a high-gradient canyon section occurs upstream of Twin Falls, while the reach upstream of Burley occupies a wide, low-gradient valley. Many of the high-gradient canyons may be associated with faults or other structural controls (O'Connor 1993).

Five major tributaries join the Snake River as it flows northward along the Idaho–Oregon border; in upstream to downstream order, these include the Owyhee, Boise, Malheur, Payette, and Weiser rivers (Figure 1.1). The drainage basins for those tributaries encompass 29,267; 10,456; 12,440; 8,614; and 4,382 km^2 (11,300; 4,037; 4,803; 3,326; and 1,693 mi^2), respectively. On the Idaho side, the Boise and Payette rivers join at RM 392 and RM 365, respectively. These tributaries drain an extensive area of high, mountainous terrain in central Idaho that includes the granitic Idaho Batholith (Othberg 1994). In these tributary basins, the geologic processes of uplifting, faulting, and glacial/fluvial activity have produced a mixture of steep canyonlands and numerous alluvial terraces (IDEQ 2000). From the Oregon side, the Owyhee and Malheur rivers join the Snake River at RM 393 and RM 368, respectively. The Owyhee River drains the Owyhee Mountains, which is an upland plateau with a metamorphic and granitic core overlain by rhyolitic and basaltic volcanic deposits that have been eroded into deep, terraced canyonlands (Mabey 1982). The Malheur River drains the southern Strawberry Mountain Wilderness, an alpine area of young volcanic rocks formed between the Basin and Range Province to the south and the Columbia Plateau to the north. The Weiser River is the last major tributary to join the Snake River (from the Idaho side at RM 351) prior to entering Hells Canyon. This tributary drains the relatively low-elevation West Mountains, which are associated with a transition zone between the Idaho Batholith and the CRBs.

The physiographic expression of the lower reaches of all five major tributaries is an important regional feature of the basin. As each of the tributaries leave their headwater areas in mountainous terrain, they flow across a more gentle, sloping plain toward the mainstem of the Snake River. For example, the average slope in the upper Boise River between the headwaters and Lucky Peak Reservoir (at the base of the Idaho Batholith) is 16 m (52 ft) per mi (0.010). In comparison, the slope for the lower Boise River between Lucky Peak Reservoir and the mainstem Snake is 4 m (13 ft) per mi (0.002). This extensive plain essentially acts as a sediment trap because the gradient is so much lower in the lower reaches (about 4 times less steep) than the adjacent highlands (this is discussed in more detail in Chapter 4). Also, these tributaries join the mainstem Snake River in a reach where the gradient drops even further, to 0.58 m (1.9 ft) per mi (0.0004), prior to entering Hells Canyon. The presence of all five tributary confluences within a relatively short distance of 40 river miles likely contributes to the abrupt geomorphologic change from the Snake River Plain to the Hells Canyon region and may be the result of structural controls (e.g., basalt flows, faulting) on stream patterns (Othberg 1994).

Downstream of Farewell Bend (RM 335), located downstream of the Weiser River confluence, the Snake River enters a deep, narrow gorge. Along this reach are the three hydropower facilities that make up the HCC: Brownlee Reservoir (RM 335 to RM 285), Oxbow Reservoir (RM 285 to RM 272), and Hells Canyon Reservoir (RM 272 to RM 247). The Burnt and Powder rivers, and smaller local tributaries such as Wildhorse River and Pine Creek, drain into the reservoirs. This area is characterized by narrow valley bottoms and mountainous terrain that rises over 914 m (3,000 ft) above the channel floor (Vallier 1998). Between Farewell Bend and Oxbow Dam the Snake River is incised into the CRB formation (O'Connor 1993).

Hells Canyon proper, in a geologic context, is commonly thought to begin below Oxbow Dam (RM 273) where the Snake River has cut through the CRBs and drains the Seven Devils (Idaho side) and Wallowa mountains (Oregon side) (Figure 1.5). These formations are resistant metamorphic and volcanic rocks that produce particularly steep canyon walls. Within the reach, the average valley elevation is approximately 1,676 m (5,500 ft) below the surrounding ridge tops (maximum of 2,438 m [8,000 ft]) (Vallier 1998), and valley walls are commonly characterized by steep talus slopes or thinly mantled bedrock. Occasional alluvial terraces are also present where the valley widens. Flow through Hells Canyon is controlled largely by these resistant landforms, as well as by ancient alluvial fans from local tributaries and historic debris flows that have altered the current river course.

Structural development and rapid downcutting have essentially precluded the development of a true floodplain and the Snake River through Hells Canyon is unusually straight with little meandering (Photo 1). Along the channel the gradient between Hells Canyon Dam (RM 247) and the Salmon River confluence (RM 188) is approximately 3 m (10 ft) per mi (0.002). The Snake River occupies almost the entire 55- to 210-m-wide (180- to 700-ft-wide) canyon and "consists of a series of straight, fault-controlled segments separated by abrupt bends" (O'Connor 1993). Notable fault zones are located at Oxbow Dam (about RM 272) and near Pittsburg Landing (about RM 215) and tend to trend northeast (Vallier 1998).

Photo 1: Hells Canyon between Bernard Creek and Sluice Creek (RM 233). Note the lack of sinuosity and floodplain, and the high Bonneville Flood terrace on the right side of the channel.

Numerous local tributaries present within the canyon drain a wide variety of basin areas, including larger drainages such as Deep Creek, Idaho (RM 247), just downstream of Hells Canyon Dam; Granite Creek (RM 240) and Sheep Creek (RM 230); and smaller drainages such as Rush Creek (RM 231) and Kirkwood Creek (RM 220). As compared to the major tributaries upstream of Weiser, these local tributaries drain smaller drainage areas. However, sediment from these tributaries is delivered directly to the mainstem from high-gradient slopes that extend directly into the river. The Imnaha River is a tributary that drains 2,200 km^2 (850 mi^2) of the Wallowa Mountains in eastern Oregon and joins the Snake River at RM 192. The Imnaha follows a fault line to its confluence with the Snake River and largely drains volcanic basalts in the CRB formation. The Salmon River, a major tributary that drains 36,260 km^2 (14,000 mi^2) in central Idaho, joins the Snake River at RM 188. This river and its tributaries flow through the extensively forested mountainous regions of the Challis, Salmon, Bitterroot, Payette, and Nez Perce national forests.

Below the confluence with the Grande Ronde River (RM 169), the Snake River valley widens somewhat and the river flows through a shallower, more gentle canyon to Lewiston, Idaho (RM 139) (Vallier 1998).

1.3. Pre-Quaternary Geologic History

Figure 1.6 presents a geologic timeline with a summary of major events. These events are discussed in chronological order in the sections below.

1.3.1. Tectonic Formation of Hells Canyon Area

The complex rock assemblages within Hells Canyon were not understood by geologists until the 1970s, despite the presence of numerous small-scale mining deposits that were developed historically. It is currently believed that many of the rocks in Hells Canyon between the oxbow located near Oxbow Dam (about RM 272) and the mouth of the Grande Ronde River (RM 169) were formed in the ancestral Pacific Ocean millions of years ago (Vallier 1998). These rocks include the Wallowa Mountains on the Oregon side and the Seven Devils Mountains on the Idaho side, and are collectively known as the Blue Mountains province. They are composed primarily of volcaniclastic and sedimentary rocks that were transported across the Pacific Ocean and docked onto the North American continent. Age dates on the mylonite suture zone between the ocean rocks and the continent show that they docked about 100 million years ago (Vallier 1998). This suture zone runs roughly parallel to Hells Canyon and is located approximately 72 km (45 mi) to the east of the canyon through the South Fork Salmon river drainage basin (Hyndman 1989).

During this docking process, the volcaniclastic and sedimentary rocks were metamorphosed under extreme temperatures and pressures. Thick sequences of basaltic and andesitic volcanic flows were metamorphosed to greenstone. Greenstones are hard, brittle, dense formations that contain plagioclase feldspar crystals in a fine-grained groundmass. Associated metamorphosed tuffs and agglomerates tend to be dark red to purple. In terms of geologic controls, the dense metamorphosed rocks comprise the steep canyon walls within the Seven Devils formation. The Seven Devils rocks tend to form massive, rugged cliffs (900 to 1,500 m [3,000 to 5,000 ft] thick) with red to maroon and dark-green to black outcrops (Miller 1975). Basalt dikes that cut through these rocks are more easily eroded and commonly form narrow, talus-filled side canyons or talus chutes (CH2M HILL 1990). Finally, small intrusive rocks are present locally within the Seven Devils formation and vary widely in age (between 115 to 300 million years old; Vallier 1998) and composition (ranging from mafic gabbros to felsic granites). One of the largest intrusions is the Wallowa Batholith, which is a diorite to grandiorite pluton that intruded into the Wallowa Mountains.

1.3.2. Emplacement of Idaho Batholith

A great expanse of pale gray granite known as the Idaho Batholith stretches across most of central Idaho (Figure 1.7). The batholith underlies an area that is at least 420 km (250 mi) long and 130 to 160 km (80 to 100 mi) wide (Lipscomb 1998). More specifically, this formation consists of two large blocks of Cretaceous granite (roughly 70 to 100 million years old), known as the Bitterroot and Atlanta Batholiths, that are separated by the Salmon River. The formation is actually comprised of large plutonic bodies with only gradational contacts; therefore, the lithology is relatively consistent. Granitic and grandioritic rocks in this formation tend to consist of potassium feldspar (K-spar) crystals intergrown with quartz, biotite mica, and hornblende. The medium-grained granite formation disaggregates easily on steep slopes and tends to erode readily under normal alpine chemical and mechanical weathering processes (IDEQ 2000). Surrounding the batholith, country rock was locally transformed into gneiss and schist as a result of contact metamorphism. The batholith is also commonly intruded by pegmatite zones and rhyolitic basalt dikes (Othberg 1994).

Following emplacement of the batholith, a portion of the overlying rocks slid along massive fault zones to the east toward Montana in the early Tertiary, exposing much of the granite body throughout central Idaho (Hyndman 1989). This is the primary reason the exposed batholith extends across such a large geographic area; normally a batholith of this size would remain covered by overlying country rocks. Despite its large areal extent, the Idaho Batholith is covered by basalt on the western margin and Snake River Plain volcanics on the southern margin (Johnson et al. 1988).

After the intrusion and subsequent faulting, Eocene volcanic deposits were erupted in a short-lived but intense episode between 40 to 55 million years ago (Moye et al. 1988). Evidence of the Eocene volcanic period is present regionally, particularly in the Challis volcanic field in central Idaho (which overlies part of the Idaho Batholith). These deposits range from andesite lava flows to rhyodacite/rhyolite ash flow tuffs and domes. The Eocene flows and ashes are largely absent in the Hells Canyon region, which suggests that they were likely eroded off of the older rocks prior to the CRB eruptions (Vallier 1998).

1.3.3. Development of Basin and Range Province

The modern Snake River Plain is located along the northern edge of what is known as the Basin and Range Province. Extending southward through Nevada and Utah, this province is characterized by northerly-tending, elongated mountain ranges that are separated from one another by broad expanses of flat basin floors. Although this crustal expansion is believed to have begun about 17 million years ago, which is approximately the same period as when the CRBs began erupting, the reasons why and the mechanisms of how this province developed continue to be studied. Eruptions from the expansion fissures and vents are believed to be the source of the thin veneer of olivine tholeiitic basalt that is present throughout the Snake River Plain (Leeman 1982, Hackett and Morgan 1988). The western Snake River Plain appears to be one of these valley floor grabens bounded by faults on both sides. As the crust in this area was thinned and pulled apart, the resulting depression was filling with upwelling basalt and interbedded debris from the adjacent highlands (Mabey 1982). Movement in this seismically-active province continues to pull the ranges apart and widen the intervening valleys along numerous normal faults, particularly in the eastern Snake River Plain. An example of an active rift area is the Great Rift zone that is partially located within the Craters of the Moon National Monument between Carey and Arco, Idaho. This feature consists of fissure vents, volcanic cones, and lava flows that began erupting only 15,000 years ago; the most recent flow in this area has been dated at 2,000 years old (Blakesley and Wright 1988).

1.3.4. Eruption of the Columbia River Basalts

During the Miocene between 14 to 17 million years ago, enormous floods of basalt erupted from fissures in the Blue and Wallowa mountains and buried pre-Cenozoic rocks to depths of thousands of feet. These basalt flows, collectively known as the CRBs, created the Columbia Plateau and flowed into the mountain valleys of eastern Oregon and western Idaho (Figure 1.8). Geologists have identified four major formations separated by periods of relative quiescence; however, of these four, only the Grande Ronde and Imnaha Formations are present within the Hells Canyon region. Basalt flows of predominantly Imnaha Formation form the uplands between Farewell Bend and Oxbow Dam; basalts associated with the Grande Ronde Formation are exposed from the Salmon River confluence to Lewiston (O'Connor 1993).

The CRBs are generally easy to recognize as massive ledges that are characterized by columnar jointing. Over 20 individual flows are present in a 700-m-thick (2,300-ft-thick) exposure near Oxbow Dam (RM 273) (Camp et al. 1982) and 35 flows have been identified in a 800-m-thick (2,600-ft-thick) exposure near the confluence of the Grande Ronde and Snake rivers (RM 167) (Reidel 1982). Imnaha basalts are classified as tholeiites with abundant and large plagioclase and small olivine phenocrysts that weather mostly to clay (Camp et al. 1982); these basalts tend to weather easily into rounded boulders and long, relatively smooth slopes as compared to the Grand Ronde basalts. In contrast, the Grande Ronde basalts are relatively aphanitic tholeiites that lack visually identifiable plagioclase and olivine; these rocks tend not to weather easily as compared to Imnaha and commonly form prominent ledges of columns with reddish soils. Basalt flows that erupted from numerous feeder vents are preserved as black and dark brown dikes throughout the canyon. Following the eruption of the CRBs, regional tectonics continued to uplift the Seven Devils and Wallowa mountains and local rivers (including the Snake, Salmon, Grande Ronde, Imnaha, and Clearwater rivers) cut deep canyons that exposed the base of the CRB flows.

1.3.5. Development of Snake River Plain

The Snake River Plain dominates the landscape in southern Idaho. The unusual shape and extent of the plain have led geologists to theorize that the plain is actually the track of a hotspot in the mantle. As the continent moved to the west over this abnormally hot, stationary portion of the mantle, the overlying crust melted and erupted to form a series of rhyolitic volcanoes. In fact, remnant caldera volcanoes in this area become progressively younger to the east (Hackett and Morgan 1988), which supports this theory. Dates on rhyolite deposits range progressively from 13 million years old west of Twin Falls to less than 1 million years old near the Wyoming border. The Yellowstone Volcano in Wyoming is believed to be the youngest member in this line of volcanoes and represents the eastern border of the Snake River Plain. This volcano is a large and active resurgent caldera that has erupted up to 600 mi^3 of rhyolite lava multiple times, most recently about 600,000 years ago (Hackett and Morgan 1988). To put this in perspective, the 1980 eruption of Mt. St. Helens erupted 0.25 to 0.50 mi^3 of material. Most of these rhyolitic formations were erupted during the later Tertiary period, and have since been buried by Quaternary olivine tholeiitic basalts that have largely originated from basin and range expansion vents (Leeman 1982).

1.3.6. Formation of Lake Idaho

Following the CRB eruptions and the development of the Snake River Plain and the Basin and Range Province, climate conditions toward the end of the Miocene (5 to 10 million years ago) appear to have changed to become more tropical with wet and warm conditions. The combination of wet conditions and large-scale volcanic activity caused a number of large regional lakes to form; as lava flows crossed rivers, the rivers were impounded and the resulting lakes captured sediments and fossils that preserve this period. For example, the downdropping Boise River Basin became a complex lacustrine sediment trap. Fossils, texture, lithology, and the structure of thick, fine-grained Miocene and Pliocene sediments indicate the presence of a lake system that captured detritus from major streams draining the surrounding highlands (Othberg 1994). Regionally, sediments in the western Snake River Plain continued to accumulate as the basin subsided through the late Tertiary. This subsidence contributed to the formation of a relatively high restricted outlet on the western margin of the plain that was likely maintained by regional volcanic drainage obstructions (Kimmel 1982). In fact, during the Pliocene, the western Snake River Plain may have drained through southeastern Oregon into northern Nevada and California (Cook and Larrison 1954, Taylor 1960, 1988; all as referenced in Othberg 1994).

The impounded water behind the restricted outlet is commonly known as Lake Idaho (Figure 1.9). Jenks and Bonnichsen (1989) concluded that a series of large permanent lakes occupied the same deep basin in the western Snake River Plain during the late Tertiary and early Quaternary; the term “Lake Idaho” generally stands for all of these episodes. Although geologic interpretation of the many terrace formations that are present in each of the major western Snake River Plain tributaries is ongoing, Othberg (1994) suggests that a complex cycle of lake transgressions and regressions occurred until the early Quaternary (approximately 1.8 million years ago). Lake Idaho sediments are characterized by basaltic and sedimentary formations that were deposited in lacustrine, fluvial, and fluvial-deltaic environments as the lake shoreline transgressed and regressed (Swirydczuk et al. 1982).

As Lake Idaho disappeared, all of the major tributaries (Boise, Owyhee, Payette, and Malheur rivers) began to deposit gravel across the broad, nearly flat western Snake River Plain. The final lake recession is marked by gravel sequences that cap lacustrine deposits throughout southern Idaho and are thought to have been deposited by braided streams (Jenks and Bonnichsen 1989). Evidence from high-elevation gravel terrace remnants throughout the valleys of the Snake River and its tributaries indicates that the tributary floodplains were at least 150 m (500 ft) higher than present levels and have since incised to their current elevations (Othberg 1994).

1.3.7. Capture of the Snake River and Incision of Hells Canyon

Between 2 to 6 million years ago, the Snake River drainage was captured and re-routed northward through Hells Canyon. Prior to this time, the Snake River is believed to have drained southwestward through the Owyhee desert, as described above (Othberg 1994). Although the mechanisms and timing continue to be studied, one theory is that as Lake Idaho episodically rose above a western volcanic divide located near the Oregon-Idaho border (the exact location continues to be studied), large volumes of water spilled over and eroded the outlet. This occurred concurrently with the southern headwater erosion of a Salmon River tributary, which eroded

Hells Canyon along structurally weakened fault zones. The southern head cutting of the Salmon River tributary may have occurred due to uplift in the Seven Devils/Wallowa area (Wheeler and Cook 1954). Eventually, the Salmon River head cutting tributary captured the Snake River near Oxbow Dam (RM 273) and diverted the flow of the Snake River northward through Hells Canyon. Erosion and capture of south-flowing tributaries near the oxbow near Oxbow Dam may explain the location and orientation of barbed tributaries in this area (e.g., Pine Creek, Wildhorse River) and the formation of the knickpoint at the oxbow (Wheeler and Cook 1954, Othberg 1994). The draining of Lake Idaho provided enough water for the Snake River to cut a deeper channel than that of the Salmon River, and consequently, the Salmon River became a tributary of the Snake River (Vallier 1998). The youngest sediments from Lake Idaho (Glenns Ferry Formation) have been dated at 2.0 to 2.5 million years old, which means the lake drained completely after that period (Vallier 1998).

The Snake River began significantly cutting Hells Canyon and flowing directly to the Columbia River around 2.0 to 2.5 million years ago. This downcutting continued not only by draining large volumes of water during glacial periods, but also by tectonic uplift along the many faults in the canyon (Figure 1.5). Normal faults appear to have uplifted large blocks of the Seven Devils Mountains hundreds to thousands of feet high (Alpha and Vallier 1994). The river appears to have preferentially eroded its course along north- to northeast-trending faults that may be associated with basin and range extension (Alpha and Vallier 1994). Vallier (1998) suggests that if faulting and earthquake activities recorded over the last 100 years occurred at a similar frequency since the Snake River was captured, it is possible that the uplift of the Wallowa and Seven Devils mountains to their present elevations was accomplished in less than 2.5 million years. During siting studies conducted for Hells Canyon Dam, bedrock in the channel was found to be overlain by up to 40 m (120 ft) of silt, sand, gravel, and boulders (CH2M HILL 1987); the presence of these materials suggests that the river extensively eroded canyon bedrock as it cut downward.

1.4. Quaternary Geologic History

1.4.1. Pleistocene Climate Conditions

The Quaternary (within the last 1.8 million years) is generally marked by a long period of glacial activity throughout the region. During Pleistocene glacial climates, the western Snake River Plain experienced cooler temperatures. As referenced in Lewis and Fosberg (1982), “on average [early] Quaternary environments were considerably different from the present; 90% of the last half a million years was more glacial than present [Emiliani 1972].” Pierce and Scott (1982) concluded that there was a relationship between periods of lowered average annual temperatures and increased gravel deposits on alluvial fans and terraces throughout the region. Textures and bedding of regional Pleistocene gravels suggest deposition by glaciated and unglaciated streams with “sustained seasonal flows probably at least ten times larger than discharges of present streams” (Pierce and Scott 1982). Relatively cooler mean annual temperatures may have caused increased seasonal discharges because of relatively thick seasonal snowpacks, and later and heavier seasonal runoff.

The influence of Pleistocene glacial activity in the Snake River Basin is a result of localized glacial activities, not the continental ice sheet. The continental ice sheet advanced only as far south as the Coeur d'Alene area and created ice dams in northern Idaho and Montana. However, active alpine glacial processes throughout the study area would have contributed large amounts of sediments to the tributaries and mainstem of the Snake River. Other factors in unglaciated areas that would increase the supply of gravel during cooler climates may include frost action that splits bedrock and solifluction that facilitates the downslope movement of gravelly material (Pierce and Scott 1982). The combination of increased discharges and increased sediment supplies caused coarse-gravel bedloads from the surrounding uplands on both the mainstem and tributaries to be transported and deposited onto the western Snake River Plain until seasonal glacial flood flows decreased. Sedimentary structures in terraces in the Boise River valley are typical of braided-stream deposits with imbricated stones and decreasing grain size in the downstream direction (Othberg 1994).

In response to these conditions, the Snake River increased scour and downcutting, particularly in Hells Canyon. In turn, multiple episodes of increased downcutting in Hells Canyon would have caused base-level lowering for the western Snake River Plain and the formation of multiple terraces in each of the major tributaries (Othberg 1994). These features were likely reworked during the catastrophic Bonneville Flood event, described in the following section.

1.4.2. Lake Bonneville Flood Event

Beginning 28,000 years ago, Lake Bonneville formed in northeastern Utah and inundated 5,200 km^2 (20,000 m^2) at its maximum extent (O'Connor 1993). Approximately 14,500 years ago, a catastrophic flood of water overtopped the outlet at Red Rocks Pass (near the Idaho-Utah border) and escaped into the Snake River Plain via the Portneuf River drainage (RM 731). The flood left large-scale geomorphic features throughout the flood route between Red Rocks Pass and Lewiston, including scabland deposits, boulder and gravel bars, and thick slackwater deposits (Malde 1968). Based on these features, O'Connor (1993) conducted a study on the hydraulics of the flood and concluded that:

- Approximately 1,200 cubic mi of water were released over a period of several weeks.
- Peak flows at Red Rocks Pass were 35 million cubic feet per second (cfs) and attenuated to 20 million cfs downstream near Lewiston.
- Stream power magnitudes ranged from less than 10 watt/m^2 in ponded reaches to greater than 100,000 watts/m^2 in major constrictions; erosion of basalt bedrock occurred primarily in reaches where the stream power magnitude exceeded 20,000 watts/m^2.

To put these calculations in perspective, the estimated peak discharge at Lewiston (20 million cfs) was 80 times larger than the current 500-year flood flow of approximately 250,000 cfs at Anatone (RM 167) located upstream of Lewiston (RM 139) (USGS 2001b).

In the upper reaches of the flood route in the Snake River Plain, the flood scoured loose rock and eroded the basalt canyon walls. Much of the debris was deposited where flow slackened into upper reaches of broader valleys. These features include melon gravel bars that are more than 30 m (100 ft) thick and extend for miles (O'Connor 1993). Channel deposits have not been significantly eroded or modified by subsequent fluvial activity (Pierce and Scott 1982). Large whirlpools and eddies in the flood waters carved deep holes into the canyon floor. In addition, gravel deposits that cover the uplands indicate that the flood overtopped the canyon walls onto the adjacent basalt flow fields.

As the flood was constricted by the inlet to Hells Canyon near Farewell Bend, a temporary lake formed and inundated adjacent valleys in Idaho and Oregon to an elevation of about 750 km (2,450 ft). Slackwater areas at the mouths of the major tributaries allowed thin-bedded clayey silt and fine sand to settle into a series of laminated beds that are typically a total of 3–6 m (10–20 ft) thick, with the thickest deposits near the Snake River channel (Othberg 1994). Slackwater deposits were likely comprised of widespread eolian silt (or loess) that was actively being deposited along the Snake River at the time of the flood (Othberg et al. 1996). The loess in southeastern Idaho appears to have originated from floodplains and active alluvial fans during the Pleistocene (Lewis and Fosberg 1982).

In Hells Canyon, depositional features are somewhat limited because of the restricted channel. Where the canyon periodically widens somewhat, large gravel terraces formed by the Bonneville Flood are present (some as high as 180 m [600 ft] above the river level; Photo 1) and into side tributaries (Vallier 1998). Flood deposits are generally subdivided into three types of bars (Webster et al. 1982):

- Point bars that contain cobbles and boulders (from 40 to 60 m [120 to 200 ft] above modern channel)
- Expansion bars that contain boulders and with finer grain sizes progressively downstream (from 40 to 60 m [120 to 200 ft] above modern channel)
- Back-eddy bars that are a mixture of gravel and sand (at least 130 m [425 ft] above modern channel).

The Bonneville Flood deposits in the canyon are generally gray to dark gray and contain an abundance of angular to subangular CRB and Seven Devils volcaniclastics (Webster et al. 1982).

Following this enormous event, the flood waters receded and the Snake River adjusted to post-flood conditions. In the reach upstream of the Owyhee River confluence (RM 393), islands within the channel formed as the slackwater deposits were incised and these relict landforms “have not been substantially altered in shape or location by recent [i.e., post-flood Holocene] geomorphic processes” (Osterkamp 1997). In contrast, the reach downstream of the Owyhee River to Farewell Bend has been “strongly affected by tributary inputs of water and sediment, and extensive evidence of post-Bonneville Flood erosion and deposition is apparent” (Osterkamp 1997). The spatial and temporal distribution of these processes is discussed in Chapter 4.

1.4.3. Holocene Climate Conditions

The Holocene period, which includes the present time, is generally considered to have begun about 10-12,000 years ago, when the last Pleistocene glacial ice age ended. Since that time, Holocene climate conditions have become gradually warmer because of the general warming trend from glacial to current regimes. However, within this general warming trend, geological records have recorded oscillating and alternating periods of temperatures that were higher than present and periods of neoglacial temperatures that mimicked previous glacial episodes (Othberg 1994). Throughout these alternating temperature cycles, Holocene floods occurred and evidence of Holocene flood events can be found in numerous hillslope (e.g., landslides, slumps) and alluvial (e.g., river terraces, alluvial fans) features in present-day Snake River Plain river channels (Pierce and Scott 1982). In contrast to Pleistocene alluvial fans and landforms in the Snake River Plain region, which consist of well-washed relatively coarse-grained gravel, Holocene sediments are predominantly fine-grained and may be reworked loess. In tributaries along the northern portion of the plain (Lost River, Lemhi, and Beaverhead ranges), fine-grained Holocene mud deposits that cover the Pleistocene gravels indicate a marked decrease in stream competency over the last 10,000 years (Pierce and Scott 1982). Based on field work conducted in this area, fine-grained Holocene alluvial fans at the mouth of streams reflect flash flooding events that are only capable of moving material much smaller than the Pleistocene gravels (Pierce and Scott 1982). This means that although evidence of Holocene flooding exists, these flood events were much smaller in magnitude than the Pleistocene floods associated with large-scale glacation.

Within Hells Canyon, higher terraces (over 30 m [100 ft] above the current channel) are believed to have been deposited by the Bonneville Flood and subsequent Holocene flooding (Vallier 1998). Some of these terraces and other Holocene alluvial fans are marked by Mazama Ash. This light white to beige layer represents an important chronological marker of wind-blown ash resulting from the eruption of Mt. Mazama (creating Crater Lake in central Oregon) 6,850 years ago. These deposits are particularly noticeable on alluvial fans where debris flows have concentrated the eroded ash from surrounding hillsides (Vallier 1994).

Lower terraces and river bars (3 to 5 m [10 to 15 ft] above the current channel) that parallel the river banks may have been formed by natural flows prior to basin regulation (more than 100 years ago). Within Hells Canyon, Holocene gravels in the river bars appear to consist primarily of sediments carried from local landslides, talus slopes, and tributary deposits. For example, Seven Devils clasts of characteristically red and green volcaniclastic lithologies are represented at Pittsburg Landing, which is located just downstream of outcrops of the Seven Devils formation (Webster et al. 1982). These deposits are discussed in more detail in Chapter 5.

1.5. Influence of Geology on Sediment Supply

The character of the Snake River and its tributaries reflects the local geologic provinces through which they flow. Although later sections of this report provide detailed analyses of specific geomorphic processes at work throughout the basin, it is important to qualitatively understand how the geologic history has affected weathering and erosion throughout the study area. More quantitative analyses are provided in Chapters 4 and 5 of this report.

Weathering is the physical and chemical breakdown of rocks, while erosion is the transport of weathered materials from the parent rock. Weathering rates are influenced primarily by rock type, exposed surface area, and underlying structure. Rocks that are indurated are more resistant to weathering. Within the Snake River Basin, the metamorphosed assemblages associated with the Owhyee Mountains are examples of these types of rocks. Larger exposed surface areas provide more places for chemical and physical weathering agents to break down the parent rocks. Rocks that are highly jointed or fractured, such as the granites of the Idaho Batholith, provide more surface area and passageways for water to enter and cause further internal weathering. Other mechanisms that increase fracturing and surface area include frost wedging (i.e., freeze-thaw cycles), heat and fire, and vegetative root growth. All of these physical weathering mechanisms are present in highest elevations of the basin, which generally coincide with the location of the Idaho Batholith. The relative weathering efficiency from these processes creates sedimentary grus (sands) material that potentially can be redistributed and transported downstream through the channel network. In addition, because the batholith contains steep slopes, eroded materials are carried more quickly downslope. This cyclic process on steep slopes exposes more fresh bedrock to the weathering and erosion process and takes less time than on gentle slopes such as those found in the eastern Snake River Plain.

In contrast to the Idaho Batholith, rocks that are being weathered and eroded within the Snake River Plain include primarily volcanic lithologies. Although basalts are generally porous, a condition that allows water to infiltrate the rock mass, the resultant chemical weathering that takes place is relatively slow in comparison to the physical processes described above. In general, the Owhyee and Malheur rivers, which drain these more resistant rock types, do not produce the voluminous sediment associated with the Boise and Payette rivers that drain the Idaho Batholith. This is consistent with other regional studies that demonstrate higher sand and fine gravel sediment yields associated with weathered granitic parent rocks, as compared to volcanic parent rocks (Lisle and Hilton 1999).

In Hells Canyon, the metamorphic rocks in the Seven Devils Mountains appear to be relatively more resistant to erosion than the CRBs. Temperature and pressures associated with regional metamorphism near the suture zone recrystallized the original volcanic and sedimentary rocks into greenstone, which tends to be relatively fine-grained, compact, and hard. Much of the overlying basalt has since been eroded, but the metamorphic rocks continue to form nearly vertical canyon walls due to fractions and joints that cut these rocks in the reach downstream of the HCC. Locally, the more resistant metamorphic lithologies stand out in contrast to weathered and eroded chutes associated with local un-metamorphased basalt dikes (Vallier 1998). Downstream of the Seven Devils formation, the Snake River and its major tributaries again erode the CRBs. Within this area, some of the minor tributaries (e.g., Granite Creek) erode dioritic plutons that are localized sources of higher sediment production. A detailed discussion of the local lithologies in relation to slopes is presented in Chapter 5.

In addition to rock types and surface area, the underlying geologic structure significantly influences erosion rates and sediment supplies; for example, the relatively straight course of the Snake River through Hells Canyon is believed to be controlled mainly by north- to northeast-trending structural faults (Photo 1; Figure 1.5). The river has cut a relatively deep canyon over a short geologic time period (less than 2.5 million years), in part, because the faults provide a zone of structural weakness that is exploited by the river's erosional energy.

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2. ANTHROPOGENIC INFLUENCES

Chapter Summary

Since the geologic framework for the Snake River system was put into place, more recent anthropogenic disturbances have impacted physical process in the study area in general, and in Hells Canyon specifically. By the 1880s land uses varied by river reach, but overall these activities likely substantially increased sediment supplies relative to pre-settlement conditions (Section 2.1.):

- Trapping—As a result of widespread trapping in the 1800s, the dwindling beaver population likely caused downstream sediment pulses as beaver dams failed.
- Mining—By the 1860s, hydraulic mining resulted in elevated sedimentation rates. For example, in Hells Canyon, placer mining activities were very common and caused the creation of new gravel bars.
- Forest Management—By the 1860s, sediment yields in timber production areas along tributaries are estimated to have increased by an order of magnitude.
- Wildfire—Prior to settlement by pioneers, Native Americans supplemented wildfires and although fire was routinely prevented by the early 1990s, between 1970 and 1995 high-intensity fires and associated erosion in the region increased dramatically.
- Agricultural Development—From 1890 to 1992, Idaho's irrigated acreage increased from 0.2 to more than 30 million acres. Livestock grazing during the late 1800s and early 1900s was unrestricted and also caused surface erosion and mass wasting in riparian zones.
- Urbanization—Within the last 40 years, the net change in sediment loading resulting from replacing agricultural land with urban land uses in the Snake River Basin has not been well quantified.

Following most of these anthropogenic effects, water storage and regulation became the most significant anthropogenic disturbance in the basin with the greatest influence on the hydrology and sediment supply (Section 2.2.). Independent from the HCC, more than 10 mainstem and 35 tributary facilities were constructed in the Snake River system between 1901 and 1969. By the time Brownlee Reservoir was completed in 1958, 87% of the upstream drainage area had already been cut off by other reservoirs, effectively keeping most of the upstream supply of sediment from entering Hells Canyon. Thus, in the last 150 years, anthropogenic disturbances first likely caused additional sediment supply to the system and then subsequent decreases in sediment supply. The combination of these factors likely produced a "slug" of sediment that has either worked, or continues to work, its way through the system.

Current land uses that continue to affect the river are rangeland, forests, cropland and pasture, and recreational uses (Section 2.3.). Recreation uses within the canyon specifically include rafting, boating, fishing, hunting, camping, and hiking.

2.1. Historical Land Uses Within the Study Area and Hells Canyon

Geomorphology and cultural/archaeology issues are closely interrelated. Specifically in Hells Canyon, researchers have integrated geomorphological studies with archaeology for many years (Cochran 1991). One paleoenvironmental study using climatic change data and information from archaeological excavations in Hells Canyon suggests that the appearance of winter villages coincides with a beginning of the period of cooler and moister conditions dating to 4200–2200 BP (Reid 1991). Recently, Langdon (2001) has demonstrated a strong correlation between Native American house pit village locations and major rapids in Hells Canyon.

Looking at the relationship *in reverse*, human activities can have landscape-level effects. For example, in Hells Canyon, historic river channel modification has affected the geomorphology. While he was directing archaeological test excavations at the Tin Shed site (RM 215), Dr. Ken Reid noticed that the river channel did not seem to have a natural thalweg (Reid, pers. comm. 1998). Reid commissioned historic archival research in order to determine whether the river channel had, in fact, been modified. The researchers determined that between 1891 and 1914, USACE blasted several rapids and removed more than 8,600 cubic yards of material in Hells Canyon in order to improve the channel for navigational purposes (Photo 2; Reid 1991).

Photo 2: Hells Canyon at Cottonwood Rapid (RM 209). Note the remnant of historical USACE dike construction upstream from the rapids.

As described in this chapter, other larger-scale historical activities that have altered regional geomorphology include water regulation, trapping, hydraulic and hard rock mining, forest management, fire, grazing, and agricultural development.

Water regulation has had the largest influence on the study area; however, this disturbance is addressed separately in Section 2.2 because the anthropogenic disturbances discussed in this section occurred prior to the construction of most of the water facilities. Aside from water regulation, significant historical anthropogenic disturbances include land uses such as trapping, mining, forest management, fire, agricultural development, and urbanization. The relative level of these activities varied by river reach and the overall influences on the study area were significant. Figure 2.1 shows the relative ranges of sediment yields of these land use disturbances in comparison to estimated yields in undisturbed watersheds (note the logarithmic scale for sediment yields). These numbers clearly suggest that many of the anthropogenic disturbances that have been documented in the study area likely caused an increase in sediment supply above pre-settlement conditions to the mainstem and its tributaries.

The timing of disturbances in the Snake River Basin is an important issue. Figure 2.2 provides a summary of the relative timing of each of these activities throughout the Snake River Basin. In general, many of these activities, such as trapping, mining, timber production, livestock grazing, and agriculture, likely caused a general increase in sediment supply above pre-settlement conditions starting in the early 1800s. As such, although some of these activities, including grazing and irrigation, continue today, the historical sediment loads produced by these anthropogenic activities began to be trapped behind the numerous reservoirs in the basin beginning in the early 1900s. Thus, in the last 150 years, anthropogenic disturbances throughout the watershed first likely caused significant additional sediment supply to the system and then subsequent decreases in sediment supply, in part, because of multiple water resource projects. The combination of these factors likely produced a “slug” of sediment that has either worked or continues to work its way through the system.

Appendix A provides a more detailed discussion of the conceptual model regarding the “slug” of sediment in the Snake River system. IPC is not maintaining that such a slug is the sole cause of changes in the sediment features in Hells Canyon, or that the HCC does not have any effect on these sediment features. IPC is maintaining that in addition to the construction and operation of the HCC Grams and Schmidt (1991 and 1999a), other larger temporal and spatial factors have played, and continue to play, a role in the dynamics of sandbars in Hells Canyon. IPC maintains that these other factors likely had an influence on sediment features (including size, shape, and quantity) in Hells Canyon that were observed and recorded by early visitors in the early to mid 20th century.

IPC is unaware of any study or watershed of comparable size and complexity in which enough data are available to be able to determine the timing and processes of sediment slug delivery to Hells Canyon, as well as to describe the geomorphic consequences of the altered sediment supply on features such as the sandbars in Hells Canyon. Even if these data were available in this watershed, Meade (1982) has recognized “for time spans such as the years, decades, and centuries over which we perceive and attempt to deal with problems of sediment movement… in rivers, we probably are least prepared to construct predictive models.”

Because there are very limited quantitative data, the timing and processes of potential sediment slug delivery to Hells Canyon are undetermined. IPC appreciates constructive reviews of the slug conceptual model; however, IPC also believes that despite the lack of direct empirical evidence, the conceptual model is valid and a large temporal and spatial sediment pulse or slug is likely, particularly from anthropogenic disturbances within the canyon itself.

Appendix A summarizes the various anthropogenic factors (e.g., trapping, hydraulic and hard rock mining, forest management, fire, grazing, and agricultural development) that have occurred throughout the Snake River watershed. Given the orders of magnitude variability in travel times documented in other watersheds and the unknown variability of travel times within the Snake River watershed, it is unclear what the relative impact from anthropogenic disturbances high in the Snake River watershed is compared to the impact from disturbances directly within Hells Canyon. Thus, the information presented in this chapter relates more directly to those anthropogenic disturbances that have been documented directly within the Hells Canyon area.

2.1.1. Trapping

Direct effects of trapping within Hells Canyon are not well documented. Regionally, the annual harvest of beaver pelts for the entire Snake River Basin (including all of its tributaries) was down to only 378 pelts by 1832 (Clark 1995). Beaver dams were historically very extensive in nearly all alluvial and low-gradient tributaries to the Snake River (NPPC 2000). They were also common in branches and backwaters of the larger tributaries and the mainstem itself (NPPC 2000). Beaver ponds effectively expand floodplains, dissipate the erosive power of floods, and act as deposition areas for sediment and nutrient-rich organic matter. The removal of beaver and subsequent deterioration of their dams typically significantly decreased floodplain storage capacity, increased peak flows, and altered sediment-supply patterns (Naiman et al. 1988 as referenced in Spence et al. 1996). Quantitative estimates of the sediment yield from beaver trapping activities have not been determined (Naiman, pers. comm. 2001), but given the widespread removal of the beaver population, the downstream sediment pulses were likely to have been significant over the 30-year trapping period.

2.1.2. Mining

Mining activities have been well documented in Hells Canyon. Alluvial mining and hard-rock mining both influenced the waterbodies in the Snake River Basin. Alluvial mining is often referred to as “placer mining” (Nelson et al. 1991), and two common types of placer mining include dredging and hydraulic mining. Dredging processes dig or suction unconsolidated alluvial deposits from beneath standing water using either large bucket-line dredges or smaller dragline dredges (Nelson et al. 1991) or motorized aquatic vacuum cleaners (Bennett 1995). All types of dredges extract the ore-bearing material from the substrate and stack the tailings away from the dredge area, typically adjacent to the channel. Hydraulic mining uses pressurized spray jets of water to extract the ore from alluvial stream banks or terraces and produce large slurries of gravel and ore. In addition to the alluvial placer mining, hard-rock mining (also known as “lode mining”) was also used within the region to extract ores from subsurface deposits. Depending on the deposit, the ore is extracted either by mechanical removal or by solution mining using chemical solvents (Nelson et al. 1991).

Mining activities within the vicinity of Hells Canyon evolved in a similar fashion as elsewhere in the study area. The discovery of gold in river bars in the 1860s near either end of the HCNRA brought a flood of human traffic (Chatters and Reid 1999). Local newspapers from the 1940s profiled Oakey Grogg, one of "hundreds of placer miners and prospectors" who worked the sediments in Hells Canyon and "contributed much of the gold that financed the Civil War."

Although rich copper lodes were discovered in 1862 and 1874 in the Seven Devils area, access to the area was difficult and interest in copper was limited at the time. However, during the 1880s and 1890s, the Seven Devils area became an active mining district. A railroad was partially built, a smelter was in operation, and three towns—Cuprum, Helena, and Landore—were established. Transporting supplies and ore proved difficult because of steep terrain; this transportation problem was the largest obstacle to profitable mining development. In 1899, gold was discovered near the mouth of the Imnaha River and more than 50 claims were taken on Squaw Creek, opposite the Seven Devils mining region. Government Land Office (GLO) records for Idaho refer to mining activity at a number of sites near the present-day HCC reservoirs between the 1870s and 1950s. Records indicate water was diverted from the Powder River for sluice mining as early as 1861 (U.S. Census Office 1896), and Chapman (1940) noted that the Powder River was generally very turbid from above Baker to its mouth because of dredge mining upstream of Baker during the 1940s. Mining was also noted at several named and unnamed locations including Brownlee Creek, Wildhorse River, Pine Creek, Deep Creek, Bernard Creek, Eureka Bar, Dug Bar, Salt Creek, Somers Creek, Temperance Creek, and Big Bar. In general, most of the placer mining in Hells Canyon occurred in the lower riparian zones; about half of these sites were flooded when the HCC reservoirs filled. Virtually all the hard-rock mining occurred higher on the slopes, generally above areas flooded by the HCC reservoirs.

When mining began, the original lone placer miner was not able to physically disturb much stream area. However, the advent of large-scale dredge and hydraulic mining activities dramatically changed the stream landscape in the basin by producing enormous volumes of sediment into stream channels and rivers. This was particularly true for south-central Idaho, where placer mining yielded more than $30 million, most of which came out of the Boise Basin (Idaho Bureau of Mines and Geology 1963). Sediment measurements on Mores Creek in the Boise River Basin indicate that placer mining operations were responsible for at least a 52% increase in sediment loads between 1939–1940 (USGS 1940). The USGS estimated that the true sediment loading increase was likely even larger because the majority of coarse sediments released by upstream placer operations were retained behind dikes and within dredge pools. Another study estimated that hydraulic mining for gold in the Boise River Basin produced 65,450 m^3 (85,600 yd^3) of silt in 18 months (Spaulding and Ogden 1968, as referenced in Nelson et al. 1991). Anecdotal evidence and newspaper articles from the late 1800s tell the story of a drowning victim that could not be found to be pulled from the Boise River because the water was too turbid from upstream hydraulic mining activities (Thornton, pers. comm. 2001). Tall piles of cobbles and dredge pools resulting from dredge mining are still evident in the Boise and Payette rivers.

In another western Idaho watershed, the volume of dredged material produced by hydraulic mining between 1860 and 1960 has been estimated at 430 m^3 (560 yd^3)/acre/year (USDA 1997). This value is based on research conducted in the South Fork Clearwater Subbasin, which is located just downstream from the study area but likely would have similar levels of mining disturbance. This sediment yield is more than 1,500 times higher than a natural erosion estimate of 0.25 m^3 (0.33 yd^3)/acre/year for debris slides in the region (USDA 1995). These region-specific values agree with other studies on the impacts of mining; in Nevada annual sediment yields in mined areas (0.57 to 7.0 yd^3 [0.44 to 5.35 m^3] /acre/year) were typically higher than in undisturbed areas (0.06 to 0.86 yd^3 [0.04 to 0.66 m^3] /acre/year) (Glancy 1973, as referenced in Nelson et al. 1991).

Within Hells Canyon, early USACE reports and more recent inventories indicate that placer mining dramatically changed the landscape (Photo 3; Harts 1899; Carrey et al. 1979). Archaeological inventories demonstrate that placer mining is represented at approximately 21% (n = 180) of the archaeological sites recorded between HCD and the Salmon River confluence (Chatters and Reid 2000). According to these data, placer mining sites cover at least 20 km (13 mi) along both riverbanks (Appendix A). This represents approximately 12% of the 167-km (104-mi) combined shoreline length of the Snake River between HCD and the Salmon River confluence. The total combined area of the placer mining sites is at least 1.52 km^2 (377 acres).

Photo 3: Picture documenting placer mining in Hells Canyon (Quicksand Creek) as late as 1963.

In addition to modifying a considerable portion of the surface area of the canyon, placer mining has also altered sediment transport in the canyon. For example, the removal of sand and gravel during placer operations leaves gravel "armor" on the riverbanks. Another example is at Salt Creek, where a recent soil column indicates that soil appears to be re-depositing following historical removal by placer mining operations. This interpretation is opposite of the general view that the soil deposits visible on beaches are remnants of larger deposits that are being washed away by riverine erosion (this issue is discussed further in Chapter 5). Other reconnaissance efforts indicate that the gravel bars near the Cochran Islands (RM 178-179) appeared to be the direct results of placer mining (Schubert 1903). Photo 4 shows Oakey Grogg and his hydraulic mining operation at Cochran Islands.

Photo 4. Oakey Grogg's hydraulic mining operation at Cochran Islands; photo taken in the 1940s.

The support systems required for mines to operate also indirectly affected sediment delivery. With dynamite and leveling equipment, miners eventually converted pack trails to the mines into roads. Such construction likely filled the streams with rocks and debris. The remnants of mine tailings are commonly found on alluvial bars and marginal sandbars within Hells Canyon (Chatters and Reid 1999).

2.1.3. Forest Management

Logging in the Pacific Northwest began in the mid 1800s, and by the 1860s the timber industry was well established (Spence et al. 1996). Within the study area, the expansion of the forest management industry was focused primarily in the central mountains of Idaho and the Blue Mountains of Oregon. Part of the timber industry grew to support the expanding mining industry, but most of the growth was due to expanding local and national markets. By 1922, nearly 1 billion board feet was offered in a single sale in the Malheur National Forest (Langston 1995). During this period, timber harvests were routinely higher than sustainable yields, partly in order to meet regional milling capacities. For example, in 1928 the USFS estimates of annual sustainable allowable cuts in the Umatilla National Forest were nine times lower (6.5 million board feet) than the milling capacity of 108 million board feet (Langston 1995). Between the 1950s and 1980s, logging activity steadily increased until the Blue Mountains were known as an "industrial forest" during the peak of production in the 1980s (Langston 1995). But by the late 1980s, a decade of drought and numerous massive outbreaks of pest infestations caused a forest health crisis regionally. Annual harvest levels on three national forests in Oregon (Wallowa Whitman, Malheur, and Umatilla) dropped from 706 million board feet in the late 1980s to less than 100 million board feet by 1993 (Langston 1995).

Forest practices have numerous effects on watershed processes. Forest practices generally include all activities associated with the access, removal, and re-establishment of forest vegetation, including road construction, timber harvest, site preparation, planting, and intermediate treatments (Spence et al. 1996). Sediment delivery from harvest, yarding, and site preparation may be increased via several mechanisms. Loss of the protective vegetative cover can increase splash erosion and decrease slope stability, while yarding activities cause extensive soil disturbance and compaction, which may increase splash erosion and channelized runoff (Spence et al. 1996). Ground-based vehicles moving logs from felled trees and skidding logs to landing sites compact and scarify the soil. In addition, compaction of the decomposing root systems reduces the infiltration capacity of these channels, leading to slumps, landslides, and surface erosion (Spence et al. 1996).

2.1.4. Wild and Prescribed Fires

For thousands of years, wildfires have affected the flora and fauna of the study area. Prior to settlement by pioneers, Native Americans also supplemented wildfires by starting their own fires. Thus, it is impossible to determine which historical fires were natural wildfires and which fires were caused by Native American communities (Langston 1995). Pioneer journals are full of entries describing conditions of smoke-filled travel through grassy areas that were burned almost every year (Langston 1995). As Native American populations declined, these burning practices ceased in the late 1800s (Clark and Sampson 1995). In the early 1900s, coincident with the rise of the timber industry, fire began to be systematically prevented (USDA 1996). However, long-term fuel conditions steadily increased such that fire frequencies and intensities currently are approaching or exceeding wildfires experienced in the early 1900s (USDA 1996). Between 1986 and 1998, large fires burned across approximately (195,162 acres) (800 km^2; 30%) of the Hells Canyon National Recreational Area (USDA 1999b). Extended drought combined with seasonal thunderstorms resulted in significant fire activity across the Intermountain West during this period.

Fire suppression was the official land management policy until the 1980s, when the role of fire as a natural disturbance necessary for the long-term health of the forest ecosystem became better understood (USDA 1996). During 1994, the study area suffered its worst wildfires in years and prescribed burning was suggested as one solution (Langston 1995). Because of the rugged terrain and predominance of flashy fuels, Hells Canyon is considered a very difficult area in which to suppress fires (USDA 1999b). An additional factor in the increased acreage tally has been a shift in fire strategies on some fires within the Hells Canyon Wilderness. Some, such as the 1994 Granite fire, were managed under a ''confine with time'' concept (USDA 1999b). The key element of this strategy permitted the fire to burn for as long as it stayed within designated wilderness and posed no threat to the public or property.

Current fires are more damaging to topsoil and produce higher erosion rates than the high-frequency, low-intensity historic wildfires (Clark and Sampson 1995). In open-canopy pine forests with a history of frequent light fires, plant roots tend to burrow deeply to escape fires and seek deeper water in the soil column (Langston 1995). In contrast, litter builds within a fire-suppressed forest, which increases the shallow soil moisture content and causes shallow roots to grow. Because shallow roots are more easily burned during fires, the resulting loss of vegetation increases the amount of soil erosion (Langston 1995). In high-intensity burns the surface soil layer can become hydrophobic, which reduces the infiltration of water and increases surface runoff (Spence et al. 1996). In addition, if fire-damaged soil is covered with heavy rain or snow melt before protective vegetation recovers, the soil more easily erodes into adjacent creeks and rivers (Clark and Sampson 1995). Thus, within the study area, a return to high-intensity fires likely is associated with an increase in overall sediment loading.

2.1.5. Livestock Grazing

Livestock grazing in the West during the late 1800s and early 1900s was essentially unrestricted from a regulatory standpoint. According to Tisdale (1986a), with the settlement of Hells Canyon, large numbers of cattle were introduced into the area’s rangelands and grazed well into the 1940s. During the 1920s, cattle grazing was mostly replaced by sheep grazing (Tisdale 1986a). By the 1940s, however, ranchers shifted back to cattle, and numerous feedlots were developed along the Snake River (Asherin and Claar 1976). Agency officials from the USDA testified to the U.S. Senate in 1936 that grazing livestock had depleted more than half the forage on western rangelands and diminished conditions on 95% of the public domain (U.S. Secretary of Agriculture 1936).

Livestock grazing can affect the riparian environment by changing, reducing, or eliminating vegetation. Riparian areas may be totally eliminated through channel widening, channel aggrading, or lowering the water table (USDA 1991). The effects of livestock grazing are especially intense in riparian zones because livestock tend to congregate in these areas (Spence et al. 1996). In riparian and upland zones, livestock increase sediment transport rates by increasing surface erosion and mass wasting (Platts 1991, Marcus et al. 1990, Heady and Child 1994 as referenced in Spence et al. 1996).

Within Hells Canyon, livestock grazing over the last century up through 1975 (when the HCNRA was adopted) has likely resulted in increased sediment loading to the Snake River and its tributaries. In Hells Canyon, riparian zones are generally narrow (less than 10 m [30 ft]) so that riparian damage is concentrated into constricted bands along creeks. Vegetation condition ratings conducted over allotments that abut the Snake River, including those within Hells Canyon, indicated that the largest portion of these areas were in only fair or poor condition (USDA 1990).

2.1.6. Agriculture and Irrigation

The Snake River corridor represents the focus of agricultural development and human settlement in Idaho. Prior to the construction of the HCC, the few flat areas and islands that existed in the upper end of the Brownlee Reservoir reach were farmed, but tillable land was limited. Flatter alluvial fans along the river and river bars were farmed by early settlers or used as feedlots for winter livestock. Later, flatter benches created by the Bonneville Flood were also farmed, and some were irrigated. Within the Oxbow and Hells Canyon Reservoir reaches, agricultural development was probably similar to that of the Brownlee Reservoir reach; river bars and benches were generally occupied by settlers raising crops and livestock. Irrigation diversions and farming are documented on the Wildhorse River and its tributaries (Arrowrock Group 1995) and probably occurred at the mouths of the few other drainages that had flatter lands that could support a house and garden or orchard. Pine Creek was reported to be dry from July through September at least during the late 1950s because of agricultural diversions (Federal Power Commission 1959).

Below HCD, the narrow canyon limited agricultural development to a few small locations that comprise less than 5% of the canyon area. Although much of the land in the canyon was privately owned, subsequent purchase of most of the land on either side of the river by the federal government also restricted settlement and agriculture. Farming and irrigation diversions have occurred at Kirkwood Ranch, Temperance Creek, Kurry Creek at Pittsburg Landing, Tin Shed, Big Bar, and Dug Bar. There was also extensive agricultural development and many irrigation diversions along the Imnaha River, which flows through mostly private lands. Farming and water diversion continue today along the Imnaha River.

2.1.7. Urbanization

Within the last 40 years, urbanization has become a more predominant land use within the Snake River Basin; however, it is important to note that urban/suburban land uses comprise less than 0.5% of the total study area. Within Hells Canyon, there is little urban or suburban development, so associated direct impacts to the Snake River are assumed to be minimal. Perhaps the most significant indirect impact from increased urbanization within the Snake River Basin is an increase in recreational visitors to the area. Because of its proximity to populated urban areas and other recreational opportunities within the HCC, this segment of the Snake River is a major recreational destination site year-round.

2.2. Water Regulation and Storage

The Snake River is currently one of the most extensively regulated and diverted rivers in North America (Palmer 1991). Consequently, historical pre-regulation flows of the Snake River and its tributaries upstream from the HCC were dramatically different than today (USBR 1998). Prior to construction of Minidoka Dam (RM 673) in 1906, irrigators diverted water from naturally flowing streams and rivers, and as farming expanded, more water was diverted to meet increasing diversion demands. Despite the construction of some small private reservoirs, streams in the basin were over-appropriated and streamflows were often fully depleted before the end of the growing season (USBR 1998). By the early 1900s, the Snake River near Blackfoot was often completely dewatered in the summer by farmers exercising their privately developed natural flow rights. For example, during the summer of 1905, the Snake River was dry for a distance of 16 km (10 mi) near Blackfoot, Idaho (USBR 1998). Periods of very low flow in the mainstem and tributaries continued to occur prior to the construction of American Falls Dam (RM 714) in 1927.

Since that time, 13 reservoirs on the mainstem and 35 facilities within its tributaries have been constructed, all of which alter the patterns and extent of river flow and sediment transport (a complete timeline of reservoir storage in the basin is presented and discussed later in this section). The smaller dams have less effect, while the larger reservoirs, such as Palisades and American Falls, more greatly influence both flow patterns and sediment passage; this is consistent with other regulated systems (Williams and Wolman 1984). Because these dams were built to allow water to be captured, stored, and diverted for irrigation, they alter the hydrograph pattern and considerably reduce the overall flow of the Snake River. For example, Figure 2.3 shows the estimated natural flow compared to the observed flow for the Snake River near Milner (RM 639) and the Boise River near Parma (RM 392). A more detailed discussion of the estimated natural flow as compared to the observed inflow to Brownlee Reservoir is provided in Chapter 4.

Along the mainstem between Jackson Dam in Wyoming and the HCC, there are currently 13 major facilities used to either store water for irrigation, flood control, or hydropower (Table 2.1). The facilities along the mainstem are operated primarily by the USBR and IPC. The first dam on the Snake River from its origin is Jackson Dam, located at Jackson Lake in Grand Teton National Park. Downstream, this dam is followed by Palisades Dam and then a sequence of weirs near Idaho Falls. Then several dams—American Falls, Minidoka, Milner, Shoshone Falls, Twin Falls, Upper Salmon Falls, Lower Salmon Falls, Bliss, C.J. Strike, and Swan Falls—lie along the Snake River prior to the HCC (Table 2.1). Together, these mainstem reservoirs have a combined storage capacity of 4.46 million acre-feet. An additional 35 major facilities (defined as having at least 5,000 acre-feet of capacity) are located along the Snake River tributaries upstream of where the Snake River flows into Brownlee Reservoir (Table 2.1). (It is important to note that there are hundreds of smaller facilities in the system that do not have at least 5,000 acre-feet of capacity.) All of the major tributary facilities (including those with less than 5,000 acre-feet of capacity) have a combined storage capacity of 5.85 million acre-feet (Table 2.1) and are operated by numerous federal, municipal, public, and privately owned entities (USBR 1998). It is important to note that Table 2.1 underestimates the total capacity because some of the values reported by sources for this table list active water capacity, not necessarily total storage capacity that would incorporate some volume for sediment retention.

Combining both the mainstem and tributary facilities, the total reservoir storage capacity in the Snake River Basin upstream from Brownlee Reservoir exceeds 10.3 million acre-feet (Table 2.1). Total reservoir storage capacity of the HCC is 1.65 million acre-feet. Thus, from a storage capacity perspective, the HCC represents less than 14% of the total storage capacity in the Snake River Basin. This percentage decreases when only the active storage capacity (1.48 million acre-feet) of the HCC is considered. Using this value, the HCC constitutes only 12.5% of the total basin storage.

The active storage component of the HCC itself can only retain approximately 11% of the average annual flow of the river (13 million acre-feet) at Weiser (USGS 2001b). This is markedly different from other large systems, such as the Grand Canyon where Lake Powell (behind the Glen Canyon Dam) alone is capable of storing 2.3 years of the Colorado River's annual flow (Collier et al. 1996).

In addition to the relative total storage capacity, the timing of the construction of the HCC is important. Figure 2.4 shows the volume of reservoir storage development in the Snake River Basin over the last 100 years. By the time Brownlee Reservoir was completed in 1958, more than 9.75 million acre-feet of storage (95% of the total storage built by that point in time and 81% of the total current storage capacity) had already been built. Prior to the completion of the Brownlee Reservoir, more than 155,900 km^2 (60,200 mi^2) of drainage area that could have potentially contributed sediment to Hells Canyon had already been cut off behind numerous other reservoirs on the mainstem and tributaries (Figure 2.5). This drainage area represents 87% of the total Snake River drainage area of 179,200 km^2 (69,200 mi^2) at Weiser. In other words, most of the drainage area of the Snake River that would otherwise contribute sediment to the system, is located upstream of dams that were constructed before 1958.

Between the Weiser River confluence (RM 351) and the Salmon River confluence (RM 188), the increase in drainage area is relatively small. From Weiser to the HCD, the drainage area increases from 179,200 to 189,850 km^2, or 10,620 km^2 (69,200 to 73,300 mi^2, or 4,100 mi^2). This drainage includes the Powder River and numerous small tributaries. From HCD to just upstream of the Salmon River, the drainage area increases by 1,400 km^2 (540 mi^2). This means that a total of 12,010 additional square kilometers (4,640 additional square miles) of drainage area contributes to the mainstem between Weiser and the Salmon River (not including the 38,460 km^2 [14,850 mi^2] from the Imnaha and Salmon Rivers). The majority of this drainage area (88%) occurs upstream of HCD.

Therefore, before construction of the HCC, the total drainage area contributing sediment to the Snake River between Weiser and the Salmon River was approximately 35,320 km^2 (13,640 mi^2) out of a total watershed area of 192,280 km^2 (74,240 mi^2) (18%). Thus, other upstream water storage projects on the Snake River and its tributaries cut off 82% of the potentially available sediment-contributing drainage area. In contrast, the HCC reduced the drainage area that contributed sediment upstream of the Salmon River by about 10,620 km^2 (4,100 mi^2). Thus, the HCC cut off only another 6% of sediment-contributing drainage area upstream from the Salmon River.

2.3. Current Land Uses Within the Study Area and Hells Canyon

Predominant land use classifications within the study area are rangeland, forest, cropland, and pasture. Table 2.2 summarizes the percentage of the Snake River Basin study area attributed to major land use classifications. Figure 2.6 shows the spatial distribution of various land use classifications under 1998 conditions.

2.3.1. Hells Canyon Land Uses

Within the HCNRA (RM 259 to RM 176), which straddles the Snake River between Idaho and Oregon and was created in 1975, the Hells Canyon Wilderness comprises nearly 870 km^2 (215,000 acres) (33%; USFS 2001). Prior to the adoption of the HCNRA in 1975, land uses included intensive local grazing and cultivation. After 1975, there was a major shift in land uses toward recreation; this shift resulted in a more stable vegetative cover with lower local erosion rates and sediment supplies. Upstream from the HCNRA in the canyon, the river corridor supports a small amount of cattle grazing. This land is considered poor quality for grazing because of the steep terrain and the overgrazed condition of the flat areas. Approximately 6 km^2 (1,500 acres) of land are cultivated within Hells Canyon above the HCNRA, based on water rights information (BPA 1985). Most of this land is around Brownlee Reservoir (BPA 1985). Private residences are widely scattered except for several small communities on the three reservoirs upstream from the HCNRA. Several summer residences are located along Brownlee Reservoir from Huntington downstream to the dam. In addition, both year-round and summer residences are located in small communities including Richland, on the Powder River arm of Brownlee; Operators' Settlement, at the base of Brownlee Dam; Copperfield, at the base of Oxbow Dam; and Homestead, 6.4 km (4 mi) south of Copperfield. These communities are all on the Oregon side (BPA 1985).

The major land use along the river corridor within the HCNRA is recreation (BPA 1995). Between HCD (RM 247) and near Pittsburg Landing (RM 214), the Snake River is designated as a Wild River, and between near Pittsburg Landing (RM 214) and RM 180 the river is designated as a Scenic River. Recreational uses include rafting, boating, and fishing and are discussed in more detail below. Some grazing and livestock forage occurs, but the terrain within the river corridor is generally not conducive for domestic animal movement or forage production. The major areas of grazing allotments are at Big Canyon, Pittsburg Landing, Sheep Creek, and Temperance Creek. Cattle and limited sheep allotments are grazed. Several ranches within the river corridor are still operated by permit at Kirby Creek, Wolf Creek, Temperance Creek, Dug Bar, Kirkwood, and Pittsburg Landing. The old ranches at Sheep Creek became a resort while the old Christmas Creek Ranch has been considered as a possible ranch base for the canyon grazing allotment. Total cultivated land within the HCNRA river section is less than 40 hectares (100 acres) (BPA 1985).

Downstream of the HCNRA, about 1 km^2 (250 acres), mostly on the Washington side of the river, are cultivated (NPS 1980 as referenced in BPA 1985). About 40 private residences are located along the 53-km (33-mi) river reach. The residences are both year-round homes and secondary or vacation homes and lot sizes are generally 0.02 km^2 (5 acres). Besides Asotin, the only other community along the 53-km (33-mi) corridor is Rogersburg, Washington, at the mouth of the Grande Ronde River (BPA 1985).

2.3.2. Hells Canyon Recreational Uses

The rugged topography and subsequent limited access to the Snake River corridor and surrounding areas controls the type and location of recreational use associated with the HCNRA. A few hardy individuals access the river corridor by foot or on horseback, but they make up a very small percentage of overall recreational use. The high gradient of the river and subsequent rapids (up to class IV) limit boat access to whitewater-type float boats and powered jet boats. Vehicle access to the river corridor is limited to four entry portals administered by the USFS.

Three of the USFS entry portals are located at the terminus of dead-end roads. Hells Canyon Creek portal is located about 1.6 km (1 mile) downstream of HCD at the end of a paved road and is about 64 km (40 miles) from the nearest sizable community (Halfway, Oregon). Pittburg Landing portal is located at the end of a well-maintained, 27-km (17-mile) long gravel road that originates at U.S. Highway 95 near Riggins, Idaho. Dug Bar portal is located on the Oregon side of the Snake River and is at the end of a 32-km- (20-mile-) long, four-wheel-drive USFS road that originates in Imnaha, Oregon, which in itself is a very isolated community. Cache Creek portal, the most northern entry into the area, is not publicly accessible by road. The portal is located at an old ranch house and is accessed by boat from several possible boat ramps located downstream.

While some hiking and hunting occurs within the HCNRA away from the immediate area of the portals, the majority of land-based recreational use occurs at or associated with the relatively small amount of easily accessible flat areas adjacent to the Snake River. These areas consist of sandbars, benches, and sandbar/bench combinations. This use is generally divided into two categories: day-users who sightsee or picnic and overnight users who use numerous designated and undesignated campsites within the area.

Boating use in the HCNRA can be divided into three distinct categories, each having its own patterns of use.

- **Commercial powerboaters**—This is the largest single use group. The majority of commercial powerboat trips are sightseeing tours with up to 40 people. These day-use trips stop at historic and cultural sites and provide a continuous interpretational commentary. A much smaller percentage of this group consists of smaller groups who offer angling or hunting. An even smaller but consistently present number of commercial powerboat customers "drop camp." They have themselves and their equipment transported to campsites and are picked up at a predetermined time.
- **Private powerboaters**—This group consists of large numbers of both day and overnight users. These users, while concentrated in the areas near portals, have the flexibility to traverse the majority of the HCNRA. Three large upstream rapids, Wild Sheep, Granite, and Rush Creek, limit many private powerboaters to either above or below this stretch. These users participate in myriad activities including angling, hunting, hiking, sightseeing, and pleasure boating.
- **Commercial and private float boaters**—These two groups are considered together because of similar use patterns. A large majority of these two groups access the river through Hells Canyon Creek portal at the upstream end of the HCNRA. With very few exceptions, these groups stay overnight while floating the river. Outfitters offer day trips, during which customers float downstream for all or a portion of a day and are transported back upstream by jet boat. Many of these users, aside from float boating, participate in angling, hiking, swimming, picnicking, and sightseeing while on these trips.

3. INVENTORY OF CURRENT PHYSICAL CONDITIONS

Chapter Summary

This chapter provides a summary of important physical conditions that affect the geomorphology of the Snake River Basin and Hells Canyon. These influences include topography, climate, geology, hydrology, soils, and vegetation. Elevations range from 224 to 3,859 m (735 to 12,662 ft), in part, because the large study area encompasses most of Idaho (Section 3.1.). Compared to Pleistocene glacial conditions, the climate appears to have gotten progressively warmer and drier for at least 12,000 years (Section 3.2.). Regional climate conditions have stabilized within the last approximately 1,000 years to a state similar to current conditions. This suggests that discharge and sediment delivery have decreased as compared to previous glacial conditions (Pierce and Scott 1982).

There are numerous lithologies within the study area because the geologic history of the region is complex (Section 3.3.). The mainstem Snake River within the Snake River Plain cuts into mafic volcanic flows (i.e., basalt), felsic pyroclastics (i.e., rhyolite), and sedimentary alluvium lithologies. On the Idaho side, the major western Snake River Plain tributaries drain the granitic Idaho Batholith, and on the Oregon side, the major tributaries cut through volcanic highland areas. Within Hells Canyon local tributaries primarily drain either basaltic volcanics or a complex assemblage of meta-volcanic and meta-sedimentary rocks that comprise the Seven Devils and Wallowa mountains. Further downstream, basalts are the dominant source rock in the mainstem and lower reaches of the Imnaha, Salmon, and Grande Ronde rivers.

The Snake River hydrograph is driven by a snow-melt regime with dynamic, flashy discharges in the tributaries (Section 3.4.). Rain-on-snow events are typically responsible for floods and debris flows, with spring runoff events accounting for the majority of annual sediment yields. More detailed hydrology discussions are presented in Chapters 4 and 5.

Soils developed on basalt bedrock in Hells Canyon are unique (Section 3.5.). Soils on the gently sloped terraces in the canyon that include a volcanic ash are productive, but are unusually highly erodable. Hells Canyon native plant communities have been affected by historic land use, exotic plant invasion, water impoundment, and management (Section 3.6.). Vegetation changes resulting from these anthropogenic factors persist over long periods of time in Hells Canyon because of the steep slopes, poor soils, and limited rainfall during the growing season (Section 3.7.).

3.1. Topography

The study area in this report includes the Snake River from its headwaters near the Idaho–Wyoming–Montana border downstream to the confluence with the Clearwater River near Lewiston, Idaho, at RM 139 (Figure 1.1). Within the study area, the drainage area associated with the Snake River covers portions of six states: Wyoming, Idaho, Nevada, Washington, Oregon, and Utah. The drainage basin includes more than 268,300 km^2 (103,600 mi^2). Figure 3.1 displays the range of elevations within the study area. Elevations range from 220 to 3,850 m (735 to 12,660 ft) above mean sea level (msl). A complete topographic summary is included in the physiographic description provided in Section 1.2.

3.2. Climate

The current climatic regime in the Snake River Basin differs from paleoclimates that have been estimated from the regional geologic record. Prior to the glacial Pleistocene period (1.8 million years ago, Figure 1.6), the global climate was typically warmer than present (Skinner and Porter 1987). Polar regions were not covered in ice and extensive tropical climates were common. Since the onset of the Pleistocene, global climate has been characterized by alternating glacial and interglacial episodes. The regional climate between 10,000 and 1,800,000 years ago may have been an average of 5 to 6 °C cooler than present conditions, based on lapse-rate calculations (Flint 1976 as referenced in Pierce and Scott 1982) and macrofossil analysis (Thompson et al. 1999). Other studies that account for precipitation gradients suggest that mean annual temperatures in the Idaho region during glacial episodes may have been as much as 10 to 15 degrees C colder than present conditions (Pierce and Scott 1982). Pierce and Scott (1982) suggest that lower temperatures likely caused thicker snowpack and delayed seasonal snowmelt, both of which would result in more rapid snowmelt and concentrated spring runoff periods with larger peak discharges than current peak discharges. (As discussed in Chapter 1, these higher discharges also resulted in greater sediment discharges with movement of relatively larger clasts.)

Since the last glacial period in the Pleistocene, Holocene geological records suggest periods with higher average temperatures than present (that is, the interval between approximately 6,000 and 8,000 years ago [Skinner and Porter 1987]). These periods were interspersed between neoglacial periods that mimicked Pleistocene glacial conditions and produced higher discharges than seen today (Othberg 1994). However, the general trend to progressively warmer and drier conditions (and lower discharges) for at least 12,000 years has been shown using evidence collected from archaeological sites within the upper Snake River Plain (Butler 1978). Global and regional climate conditions during the last 1,000 years are believed to have been relatively similar to the present regime (Skinner and Porter 1987, Hydrocomp 1990). Regionally within the Snake River Plain, Pearson (1978) documented periodic drought periods using tree ring growth records over the last 900 years (Figure 3.2). These results suggest that the Snake River area followed other regional trends because major drought periods in the region are similar to tree ring growth patterns elsewhere in the western United States (Hydrocomp 1990). In general, the trends shown in Figure 3.2 suggest that the regional climate has fluctuated around, and been similar to, current conditions since 1100 AD.

In summary, although Pleistocene glacial conditions produced significantly higher discharges than current discharges, Holocene climate conditions have oscillated between warmer and neoglacial periods. The resulting cyclic nature of discharges appears to have stabilized within the last 1,000 years to a state similar to current conditions. As such, the hydrologic and sediment transport processes at work in the basin have also been relatively stable in the last approximately 1,000 years; throughout this period, discharge and sediment delivery have decreased as compared to Pleistocene and Holocene glacial conditions (Pierce and Scott 1982).

The current climate in the Snake River Basin is influenced primarily by Pacific maritime polar air masses that travel eastward over the continent (Abramovich et al. 1998). The basin tends to have warm, dry summers from the western desert continental climate and cold, moist winters from prevailing westerly coastal storms (Abramovich et al. 1998). Temperatures vary with elevation (Figure 3.3). Mean annual temperatures in the valleys and regional lowlands (such as the Snake River Plain) are about 10 °C. At higher elevations, average temperatures are about 2 °C (WRCC 2001). Minimum temperatures less than minus 40 °C occur in the mountains and maximum temperatures greater than 43 °C occur in the lowlands.

Annual precipitation extremes are also common between the valley and mountain areas because of significant elevation differences (Figure 3.4). The normal annual precipitation ranges from less than 25 cm (10 inches) in the western and eastern Snake River Plain to more than 127 cm (50 inches) in the Sawtooth Mountains located north of the plain (University of Idaho 1995). In most areas of the basin, the highest monthly precipitation occurs in December and January. Monthly totals decrease irregularly until July and August, which are the driest months (Abramovich et al. 1998). The heaviest precipitation occurs in the Sawtooth Mountains, where a snowpack of several feet accumulates and remains year-round in some areas (Lipscomb 1998). For those mountainous areas in central Idaho that receive 76 cm (30 inches) or more of annual precipitation, 59% drain to the Salmon River to the north and 41% drain to the tributaries along the Snake River Plain to the south (e.g., Payette, Boise, Wood, and Lost River watersheds). Snowline elevations in the basin range between 1,100 to 2,700 m (3,500 and 9,000 ft), depending on the annual climate patterns and topographic influences (WRCC 2001). The average annual snowfall depths range from less than 1.0 foot around Lewiston to about 3 m (10 ft) along the Idaho-Wyoming border (Abramovich et al. 1998).

Hells Canyon proper is significantly influenced by the rain shadow of the mountain ranges to the west. The average annual precipitation for Brownlee Dam and Lewiston, Idaho weather stations ranges from about 30 to 46 cm (12 to 18 inches). Nearly 45% of the average annual precipitation at the Brownlee Dam weather station falls between November and January, which contrasts strongly with the 9% that falls between July and September (Johnson and Simon 1987). Precipitation is bimodal with intense, short duration summer storms and milder, longer duration winter storms (Abramovich et al. 1998). Much of the water in this reach is derived from snowmelt runoff from high elevations and upstream reaches of the mainstem and the tributaries.

Mean annual temperatures are similar among the two NOAA weather stations. Generally, the climate becomes drier and warmer downstream of Brownlee Reservoir. The canyon bottom area is dry with seasonal temperatures ranging from about minus 5 degrees C in January to about 35 degrees°C in July. At elevations above 1,000 m (3,280 ft), mean temperatures range from 0 degrees°C in January to between 28 degrees°C and 33 degrees°C in July (Johnson and Simon 1987). As a general rule, winters in Hells Canyon are relatively mild, while summers are hot.

In terms of future climate change, although the Snake River climate is dynamic within a geologic time frame (Pierce and Scott 1982), records over the last approximately 1,000 years show a reasonably stable climate (Pearson 1978, Butler 1978) that would not be expected to change significantly over the next 30 to 50 years.

3.3. Geology

There are numerous lithologies that are present within the basin because the geologic history of the region is complex (Figure 3.5; Chapter 1). Table 3.1 summaries the dominant lithologies for each of the major reaches and tributaries of the Snake River. Although specific channel bed lithologies are not well documented over the large study area, channel sediment composition likely reflects the surrounding lithologies.

In general, the path of the mainstem Snake River within the Snake River Plain cuts into mafic volcanic flows (i.e., basalt), felsic pyroclastics (i.e., rhyolite), and sedimentary alluvium lithologies. Many of the small tributaries in the eastern Snake River Plain have their headwaters in sedimentary rocks; however, these tributaries do not represent a significant source of water or sediment to the mainstem (Clark et al. 1998). One the main reasons for this is that these tributaries discharge into the underlying porous volcanic materials, such that much of the runoff does not reach the Snake River through surface water pathways. An extreme example of this process is the Big Lost River, which disappears entirely into the underlying rocks prior to reaching the Snake River. This water becomes groundwater that travels through the subsurface basalt aquifer and eventually seeps through the canyon walls into the mainstem Snake River below Thousand Springs.

The major tributaries in the western Snake River Plain drain alkalic intrusives (i.e., the headwaters of the Boise and Payette rivers in the Idaho Batholith), felsic pyroclastics (i.e., the Owyhee and Malheur cut through volcanic highland areas), and/or mafic volcanic flows (i.e., basalts in the Weiser River basin). Near the lower reaches and the mouths of each of the western Snake River Plain tributaries, sedimentary and metamorphosed sedimentary rocks are also present (i.e., Lake Idaho deposits).

Downstream of Farewell Bend, local tributaries to the Hells Canyon Complex primarily drain either basaltic volcanics (e.g., CRB flows within the Burnt and Powder river basins) or a complex assemblage of meta-sedimentary rocks and pockets of alkalic and mafic intrusives (e.g., granite and gabbro plutons). Downstream of Hells Canyon Dam, local tributaries begin to drain more meta-volcanic and meta-sedimentary rocks that comprise the Seven Devils and Wallowa mountains (Figure 1.5). Isolated diorite intrusives are present, but they are not aerially extensive. Further downstream, the mainstem again begins to cut CRB mafic volcanic flows that are at the river level near Pittsburg Landing at RM 215. These basalts are also the dominant source rock in the lower reaches of the Imnaha, Salmon, and Grande Ronde rivers. Finally, the

Snake River flows through loess-covered CRB flows between the confluence of the Grande Ronde and Lewiston, Idaho.

3.4. Hydrology

The Snake River hydrograph is influenced by a snowmelt regime. Headwater streams are generally fed by snowmelt with peak flows in the spring. For example, Osterkamp (1997) states that annual hydrographs of the major tributaries upstream from Weiser "are influenced strongly by spring snowmelt, but, relative to that of the upper Snake Basin, geologic conditions in the tributary basins result in faster runoff and flow durations that cover two to three orders of magnitude. For this reason, and because mean elevations of some tributary basins are lower...peak flows of some tributaries ordinarily precede those of the Snake River." In contrast, the mainstem river at lower elevations tends to be a non-flashy, groundwater-dominated discharge with annual flows attenuated by surface water and groundwater interactions (Figure 3.6; USBR 1998). These observations are consistent with common hydrologic principals whereby discharges in smaller drainages tend to be more dynamic and flashy than discharges in larger catchments.

Peak runoff in many of the major and minor tributaries to the Snake River can be influenced by rain-on-snow events, where warm spring rains cause rapid melting of the mountain snow cover and substantially augmented discharges. The wide-spread flooding and tributary debris flows recorded in the winter of 1997 were a result of such an event. For example, following this storm, 483 landslides were observed in an inventory area of 760 km^2 (294 mi^2) in the west side of the Payette National Forest (Dixon and Wasniewski 1998). This landslide density is significantly higher than natural rates, where only a handful of scars from landslides occurring between extreme rain-on-snow events are evident (Dixon, pers. comm. 2001). Of the 483 landslides associated with the 1997 storm event, 66% created debris torrents that essentially scoured the channels that were downslope from the landslide (Dixon and Wasniowiski 1998). Even in normal precipitation years, spring runoff events may account for up to 90% of the annual sediment yield in first- and second-order watersheds in central Idaho (Megahan 1982 as referenced in Johnejack and Megahan 1991).

Because the current hydrological conditions play a significant role in the geomorphology of the Snake River and Hells Canyon, stream flows are described in more detail in Chapter 4 (upstream from Hells Canyon Dam) and Chapter 5 (downstream of Hells Canyon Dam). These chapters also discuss unregulated conditions to provide a comparison against current conditions.

3.5. Soils

The formation of soil is influenced, in part, by parent rock material that has been altered over time by climate, location, and biological organisms. The complex lithology of the region has created a somewhat complex soil coverage. Because the Snake River Basin covers such a large spatial extent, only dominant soil types are summarized.

Throughout the basin, dominant soil types range from aridisols on the Snake River Plain to inceptisols in the central mountain region of the Idaho Batholith (Figure 3.7). Aridisols in the Snake River Plain, which are a result of the semiarid climate, tend to be low in organic matter and are typically light in color. Aridisols support sagebrush and bunchgrass and are used mainly for farming, as well as for rangeland habitat (SCS 1980a). These soils are characterized as shallow to very deep, and are somewhat poorly drained to somewhat excessively drained (SCS 1980a). A thin veneer of loess and windblown sand has generally been incorporated into the soil profile throughout the plain. These loess deposits are thickest near the Snake River and decrease in the thickness away from the river (Othberg 1994). Besides being highly fertile, loess soils tend to absorb and hold water beyond one growing season and are used heavily for agriculture.

On terraces and uplands in the plain, the downward migration of water through the soil profile has transferred and deposited soluble salts and carbonates to form calcium- and silica-cemented hardpans—duripans—in the lower horizons (SCS 1980b). Within the western Snake River Plain, the amount of soil development on the terraces is generally proportional to age. Older terraces typically have stronger duripan development and thicker clay horizons (Othberg 1994). Terrace aridisols range from well to excessively drained, and are suitable for pasture and agronomic uses, and wildlife habitat (SCS 1980b).

The foothill areas that are located to the south and north of the plain are typically covered by mollisols. Mollisols are well-developed soils that tend to have organic-rich surface horizons and clay-enriched lower horizons. These soils typically support forested vegetation, as well as rangeland and wildlife habitat, and are characterized by moderately deep to very deep with well-drained properties (SCS 1980a). Entosols that dominate throughout the southern areas of the Idaho Batholith are relatively young and lack soil horizons (USDA 1998). This is due primarily to colder temperatures in the higher altitudes, relatively steep and actively eroding slopes, and inertness of the parent materials (quartz-rich granite) (USDA 1999a). Central areas of the Idaho Batholith are covered more extensively by inceptisols, which are characterized by weakly developed soil horizons. Inceptisols commonly occur on relatively active landscapes, such as mountain slopes where active erosional processes expose unweathered materials and fluvial or glacial valleys where relatively unweathered sediments are deposited (USDA 1999a).

Within Hells Canyon, soils are derived primarily from CRBs and are either mollisols or aridisols (Jones 1993). Soils developed on basalt bedrock in Hells Canyon are unique because the layered basalt flows have produced horizontal bands of shallow to very deep soil, interrupted by bands of scabland or minor cliffs (USDA 2000). Most rock faces are nearly vertical, with little soil cover. The dominant soil complexes are well drained and vary from very shallow to moderately deep. Residual soils are frequently deep enough on north-facing canyon slopes to support conifers; however, grassland prevails on generally shallow soils on south-facing canyon slopes (USDA 2000).

Xerolls, the most common mollisol suborder, are developed in a moist winter, dry summer climate and are continually dry for long periods of time. These soils dominate the steppe and shrub-steppe vegetation areas in the canyon. The two most common aridisols in this canyon are those with accumulations of calcium carbonate and other salts (calcids), and those that are distinguished by the accumulation of clay in subsurface horizons (argids) similar to the Snake River Plain upstream from the HCC. Soils on the gently sloped terraces in the canyon that include a volcanic ash cap (for example, Mt. Mazama ash) are productive, but are highly erodable (USDA 2000). Erosion has removed the ash on steeper slopes and stream bottoms.

3.6. Vegetation

Vegetation in the study area varies widely. Within Hells Canyon, native plant communities have been affected by several anthropogenic factors. Figure 3.8 shows the change in vegetation cover between pre-settlement conditions and current conditions and is classified in terms of hydrologic cover, which provides a relative measure of how well the vegetation and litter/duff cover protects the soil from erosion. As shown in the figure, the decrease in very high and high vegetation hydrologic cover reflects a number of anthropogenic and ecological factors. These factors are historic land use, exotic plant invasion, and water impoundment and management. Changes resulting from these anthropogenic factors persist over long periods in Hells Canyon because of the steep slopes, poor soils, and limited rainfall during the growing season. Of the three ecological considerations, climate exerts the strongest influence on the development of plant life. As described in Section 3.2., the region is semi-arid, falling in the rain shadow of the Cascade Range to the west. Within Hells Canyon, for example, the relatively mild winters below the canyon rim have allowed the development of disjunct species such as hackberry (*Celtis reticulata*), a native species that is most often found in the southwestern states, but commonly occurs in the Middle Snake River area (Tisdale 1979).

Within the Hells Canyon reach, a narrow band of diverse riparian communities follows the course of the Snake River and its tributaries. Although it is limited in geographic area, this riparian zone is vital because of the biological diversity it provides (BLM 1986). Predominant tree species in riparian areas include silver maple (*Acer saccharinum*), European willow (*Salix* X *rubens*), white alder (*Alnus rhombifolia*), water birch (*Betula occidentalis*), and, along some tributaries, black cottonwood (*Populus trichocarpa*). Predominant shrub species in riparian areas include syringa (*Philadelphus lewisii*), netleaf hackberry (*Celtis reticulata*), chokecherry (*Prunus virginiana*), black hawthorn (*Crataegus douglasii*), and poison ivy (*Toxicodendron radicans*). Native willow communities are rare along the mainstem river.

Riparian communities are most prominent above Brownlee Reservoir and along Oxbow and Hells Canyon reservoirs and sparse and scattered along Brownlee Reservoir and below Hells Canyon Dam. Many exotic species are present in the riparian zone above Brownlee Reservoir while a mix of native and exotic species occur along Oxbow and Hells Canyon reservoirs. Although no reclamation activities have been undertaken with the Hells Canyon National Recreation Area (HCNRA) (RM 259-RM 176), riparian vegetation along tributaries has recovered somewhat from past grazing and mining abuse because of the changes in land use after the HCNRA was designated. Riparian vegetation along tributaries outside the HCNRA has not recovered as well because of continued livestock grazing and water diversion.

At many locations, upland vegetation on steep canyon slopes simply meets a rocky shoreline. Where this situation occurs, the dominant species are bluebunch wheatgrass (*Pseudoroegnaria spicata*), cheatgrass (*Bromus tectorum*), and Idaho fescue (*Festuca idahoensis*) (Asherin and Claar 1976). Although coniferous forest communities are generally restricted to the higher elevations of steep canyon slopes, they do reach down as far as the river at certain locations. This is the case at sites around the main bodies of Oxbow and Hells Canyon reservoirs, where a ponderosa pine (*Pinus ponderosa*)/bluebunch wheatgrass type extends to the river on north-facing slopes (Asherin and Claar 1976, BPA 1985). A ponderosa pine/hackberry type may also extend down to the river in these areas.

Within the context of regional climate, topography is a major influence on the development and distribution of vegetation (Tisdale et al. 1969, Tisdale 1979, Tisdale 1986b). Tisdale (1979) noted that variations in aspect, elevation, and slope gradient often cause dramatic differences in microclimate, thereby affecting the character of both soils and vegetation. As a result, the topographical complexity of Hells Canyon has produced a mosaic of vegetation types (Tisdale 1979, BPA 1983, BLM 1986). Grassland, shrubland, riparian, and coniferous forest communities exist in proximity. Inter-fingering of grassland and forest, for example, occurs at a number of sites throughout the canyons as a result of these variations in aspect (Tisdale 1979).

No detailed knowledge of the character of wetland and riparian communities in Hells Canyon is available. Descriptions are based on work that focused on particular plant communities (Miller 1976, Miller and Johnson 1976, DeBolt 1992), or is based on limited sampling efforts (Huschle 1975, Asherin and Claar 1976). Emergent wetland communities in the study area are composed mostly of common cattail (*Typha latifolia*), narrowleaf cattail (*Typha angustifolia*), American bulrush (*Scirpus americanus*), and common spikerush (*Eleocharis palustris*). A common cattail/American bulrush community type occupies shallow shorelines, shallow bays, and ponds. The plants grow to a height of approximately 2 m (6.5 ft) and their distribution varies in density. Willows are sparsely represented, and various forbs grow on the shoreline side of the stands (Asherin and Claar 1976). A common cattail/common spikerush community type occurs on some tributary deltas. Narrowleaf cattail is present, though sparsely distributed. Willows are slightly more abundant than the common cattail/American bulrush type (Asherin and Claar 1976).

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4. Hydrology, Morphology, and Sediment Dynamics Upstream from Hells Canyon Dam

Chapter Summary

Stream flows and sediment dynamics are the primary influences on geomorphology in the Snake River system. These processes upstream from HCD are described in this chapter, while the interaction between stream flows, morphology, and sediment dynamics in Hells Canyon below the HCC is described in Chapter 5.

Water and sediment that enters into Brownlee Reservoir is directly related to the processes at work in the mainstem reach immediately above the HCC, which in turn is influenced by inputs from the next upstream mainstem reach and the Owyhee, Boise, Malheur, Payette, and Weiser rivers (Section 4.1.). Prior to the completion of Brownlee Reservoir in 1958, more than 10 upstream facilities on the mainstem and 35 tributary water storage projects were already in place. These facilities have had an extensive impact because together they are estimated to have trapped 92% of the available sediment from 1901 through 1999 upstream from Brownlee Reservoir.

These facilities have also altered stream flows in this portion of the study area because so much of the water is either stored or diverted. During dry years, almost half of the estimated naturally occurring volume of the Snake River is diverted for agricultural purposes and during wet years approximately one-third is diverted. Most importantly from a geomorphic context, these projects have also reduced peak flow rates that transport the majority of sediment load. Thus, the available sediment transport capacity has continued to decline compared to historical (geologic) conditions.

This reach of the Snake River is located in a relatively flat alluvial plain, where coarse sediment that reaches the mainstem is generally deposited prior to reaching the HCC. A combination of upstream regulation projects and a flat gradient have caused channel islands upstream from the HCC to increase in areal extent by an average of 8% since 1938. Thus, the mainstem and tributary storage projects have reduced peak flows and decreased the volume of fine and coarse sediment transported downstream past Weiser compared to pre-regulation conditions.

Sediment that does reach the HCC from the upstream reach is composed primarily of silt-clay and very fine sand sediments (97%; Section 4.2.). In the absence of the HCC, fine-grained silt-clay sediments would flush through the Hells Canyon reach quickly. Coarser sand and gravels (3%) are present primarily in the headwaters of Brownlee Reservoir. Local tributaries also produce sediment that is retained in the HCC and this material contains a relatively higher proportion of sands and gravels. The sands and coarser sediments that have been retained upstream as a result of the HCC are only a small fraction of pre-regulation sediment supplies in the basin.

Hillslope and fluvial geomorphology reflect a continuum of process, whereby local hillslope processes deliver sediment and organic material to stream channels, and fluvial processes transport and deposit this material downstream through the channel network. In the Snake River system, stream flows and sediment dynamics are the primary physical processes that affect the geomorphology of the river. Each of these components is discussed in detail in the following chapter, which covers the portion of the study area that extends from the Snake River headwaters to the HCD (Chapter 5 contains a similar evaluation for Hells Canyon below HCD).

IPC understands that a sediment budget would be a useful tool for assessing the relative effects from the HCC on downstream resources. To support this report, the development of a sediment budget was evaluated for the portion of the basin upstream from the HCC. Typically, a sediment budget defines the various sediment sources within a basin, including consideration of the available storage, and routes this sediment to and through the channel system (Reid and Dunne 1996, Dietrich et al. 1982). In an ideal evaluation, pre- and post-regulation sediment budgets would be compared to help evaluate the specific impacts of the HCC on sediment delivery to Hells Canyon. However, in a study area as large as this (more than 179,200 km^2 [69,200 mi^2] upstream from the HCC) with numerous anthropogenic disturbances (some of which are related to regulation, some of which are not), the subsequent range of sediment processes is extremely complex to quantify. In larger watersheds, sediment budgets are typically difficult to define. For example, even in the Grand Canyon where extensive research has been conducted for more than 20 years, “one elusive goal of Grand Canyon researchers has been to provide managers an accurate budget of inflows, outflows, and changes in storage of sand” (Schmidt 1999). Thus, even in situations where data are relatively comprehensive, sediment budgets may only be able to provide a conceptual framework.

Nonetheless, since the draft report was produced, a sediment budget was developed based on the limited data that are available; the sediment budget by size class is presented in Section 10.7. in Parkinson et al. (2003). The results of the estimated budget are presented in this chapter for those components upstream from HCD, and the results for those components downstream of HCD within Hells Canyon are presented in Chapter 5.

In addition to the sediment budget, an estimate can be made of the overall sediment yield that the basin has produced over the last 100 years. This estimate provides insight into the relative magnitude of sediment delivery into the HCC from this upstream area and places the relative effects from the HCC into the context with the other 45 upstream regulation projects on the Snake River system. In addition, the overall sediment yield into the HCC captures the integrated effects of physical processes throughout the basin.

With the development of large-scale storage facilities beginning in the Snake River Basin in 1901, sediment that was once available for downstream transport began to be trapped (Chapter 2). Based on the number and timing of storage projects and the contributing drainage areas and sediment yields, the quantity of sediment that would have been available to the Brownlee Dam location from 1901 through 1999 in the absence of basin-wide storage has been estimated.

Figure 4.1 is a schematic diagram that shows the reservoirs and sediment yield values included in the computation (Appendix B). A total of nine sediment yield values based on reservoir sedimentation surveys were available for major six reservoirs in the basin (USBR 1999). A sediment yield of 0.15 acre-feet/m^2/year (0.38 cubic yards/acre/year) was used as an estimate for those reservoirs for which published values were not available.[1] Although not all storage facilities within the Snake River Basin are shown in Figure 4.1, their drainage areas are accounted for in the calculation.

Based on the method described above, approximately 838,400 acre-feet (1,826 million tons) of sediment have accumulated in mainstem and tributary storage facilities upstream of Brownlee Reservoir from 1901 through 1999. Similarly, using 0.15 acre-feet/m^2/year for sediment yield to Brownlee Reservoir, approximately 68,600 acre-feet of sediment have accumulated within the reservoir since storage began in 1958. This estimate compares reasonably well with the measured sediment volume of 62,046 acre-feet from the 1999 bathymetric survey of Brownlee Reservoir (Butler 2001).

Thus, the amount of sediment accumulated within Brownlee Reservoir represents approximately 8% of the total volume of sediment accumulated within the basin upstream of Brownlee Dam. In other words, 92% of the available sediment from 1901 through 1999 has already been trapped upstream from Brownlee Reservoir. On the basis of annual sediment yield, approximately 15% of the pre-1901 annual sediment yield is now available for transport to Brownlee Reservoir. Most of this material consists of fine-grained sediments that are transported in suspension and have a relatively minor effect on river channel form. This computation reveals the impact of the extensive storage upstream of the HCC.

A more detailed discussion of the physical processes that contribute to this sediment yield, as well as a characterization of the sediment that does reach the HCC, is provided in the following sections. This chapter is essentially divided into two major geographic areas: upstream from the HCC (Section 4.1.) and within the HCC (Section 4.2.).

4.1. Upstream from the HCC

The spatial extent of this portion of the study area upstream from the HCC contains more than 67% of the total study area of 268,300 km^2 (103,600 mi^2). Thus, stream flows, morphology, and sediment dynamics are divided into three different sub-geographic areas: [1) the mainstem Snake River upstream from the Owyhee River, 2) the major tributaries between the Owyhee and Weiser rivers, and 3) the mainstem between the Owyhee River and Brownlee Reservoir.] These three geographic areas are divided as such because what enters into (and is retained by) the HCC is directly related to the processes at work in the next upstream mainstem reach. In order to put this mainstem reach into the proper context, it is necessary to also understand system inputs from the majority tributaries and from the next upstream mainstem reach. These three geographic areas are discussed in the downstream direction in the following section.

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[1] To determine a basin-wide average value of 0.15 acre-feet/m^2/year, the reservoirs with the relatively lowest yields in the basin were averaged. (If we had used every reservoir value, including the higher-yield Cascade Reservoir that drains a relatively small watershed, the basin-wide average would have been 0.23 acre-feet/m^2/year.) Although using such a generalized approach introduces potential for error into the estimate, the fact that the estimated watershed-wide yield compares well with the measured sediment volume from the 1999 bathymetric survey of Brownlee Reservoir suggests that 0.15 acre-feet/m^2/year is a reasonable approximation of upstream sediment retention by other regulation projects.

4.1.1. Mainstem Upstream from Owyhee River (RM 393)

4.1.1.1. Stream Flows

This section describes the current stream flows throughout the Snake River Basin upstream from the HCC. These flows reflect significant anthropogenic alterations such as water storage reservoirs and irrigation diversions. A comparison between current regulated conditions and estimated unregulated conditions is provided where available.

Average annual river flow of the mainstem Snake River and various tributaries varies widely by location, primarily as a result of tributary inflows, diversions, and interactions with the underlying aquifers. Figure 4.2 shows the mean annual flow[2] throughout the Snake River Basin between Heise, Idaho, (RM 854, upstream from the Henrys Fork confluence) and the Owyhee River (RM 393; USGS 2001b). (This figure includes discharges through the lower extent of the study areas to Anatone, Washington, [RM 167] for comparison purposes). In the uppermost reaches of the Snake River, the river flows through a fertile plain that is heavily irrigated (USACE 1999). The seasonal flows result mostly from natural snowmelt that has been modified by reservoir storage operations (USACE 1999). At Heise, which is located upstream from nearly all irrigation uses, the average annual flow is about 7,000 cfs (USGS 2001b). Downstream of Heise (RM 854), a significant amount of the river flow is lost to groundwater and naturally recharges the eastern Snake River Plain Aquifer (USACE 1999). Henrys Fork and other major tributaries provide an additional average flow of 3,100 cfs within this section of the mainstem (IDEQ 1997). However, by the time the river passes by Milner Dam (RM 639), annual flows decrease to an average flow of 2,850 cfs (USGS 2001b) because of irrigation diversions and groundwater recharge. In the driest years, average summer flows at Milner Dam may be less than 700 cfs and may occasionally be reduced to zero (IDEQ 1997).

Downstream of Milner Dam, flows increase substantially from groundwater discharge, irrigation returns, and some smaller tributaries. At Thousand Springs near Hagerman, Idaho, (RM 572), the eastern Snake River Plain Aquifer discharges into the mainstem canyon. In the late 1990s, groundwater provided an average annual inflow of about 5,300 cfs to the mainstem (USACE 1999; IDEQ 1997). This discharge used to be as high as 6,800 cfs in the 1950s, when groundwater levels were higher due to deep percolation of water applied to irrigated lands, seepage from conveyance canals, and the decreased use of groundwater for drinking water (USACE 1999; IDWR 1999). Tributaries in this reach contribute an additional average flow of 2,300 cfs to the mainstem (IDEQ 1997). Finally, more than 80 agricultural drains return water to the mainstem between Milner Dam and King Hill; the return flows are highly variable and contain significant portions of groundwater. Between King Hill (RM 547) and the Owyhee River confluence (RM 393), flows in the mainstem increase primarily from tributary inflows from the

[2] For all of the gages on the mainstem upstream from the HCC, the period between water years 1926 and 2000 (October 1925 to September 2000) was used to represent post-regulation conditions because American Falls Reservoir was completed in 1926. For mainstem gages below HCC, the available post-construction period is used as the period of record. For all other tributaries, available data from the period following the most recent major storage facility (Table 2.1) are used as the period of record. Mean annual discharge was calculated as the arithmetic average of mean water-year discharge values. Mean minimum and maximum discharges for each station are also included to provide an indication of possible ranges of discharge throughout the study area. The mean minimum and maximum discharge was calculated as the arithmetic average of minimum and maximum water-year discharge values, respectively.

Bruneau River (RM 495) into C.J. Strike Reservoir and irrigation return flows/runoff (IDEQ 1997).

Current flows are different from both historical (geologic) and pre-regulation conditions. Pre-regulated peak flows are estimated to have been up to 10 times smaller than the historical discharges that presumably formed the existing channel during the Pleistocene and Holocene (Pierce and Scott 1982). Following the marked decrease from historical (geologic) conditions, numerous storage reservoirs and irrigation diversions on the Snake River and its tributaries also altered the delivery and transport of water and sediment. Estimated pre-regulated flows and regulated flows at Milner, Idaho, were previously compared in Figure 2.3.

On the mainstem Snake River, Osterkamp (1997) believes that “regulation of the Snake River by dam releases and reduction of flow by groundwater and surface water extractions have altered pre-development discharge characteristics.” Specifically, at the Murphy, Idaho, (RM 453) gage mean daily discharge under pre-regulated conditions (pre-1926[3]) varied from “about 5,400 to 45,000 cfs. Under regulated conditions (post-1926), mean daily discharge has varied even less, from about 4,400 to 35,000 cfs” (Osterkamp 1997). The pre-regulation range of flows is relatively small as “most streams vary in discharge within three to four orders of magnitude” (Osterkamp 1997), while the post-regulation range of flows is even smaller. This suggests that peak flows, which typically transport the majority of fine and coarse sediment (Leopold et al. 1964), have continued to decline compared to historical (geologic) conditions because of changes in regulation and development.

4.1.1.2. Morphology

Classification and characterization schemes of rivers based on morphology and processes are numerous. Most watershed and river systems, including the Snake River, fit the classic description of the “idealized fluvial system” developed by Schumm (1977) to better understand the key geomorphic zones (Figure 4.3). This system consists of three zones as follows:

- The classic idealized fluvial system originates in an upper watershed area where most of the water is produced. This area generally has significant topographic relief consisting of mountains and valleys. In addition to producing water in the form of surface runoff, this area also produces sediment. These fast mountainous streams have high erosive energy and can transport considerable amounts of sediment through this zone, which has been called the production zone.
- After flowing from the upper watershed area, the classic river system flows through a relatively flat alluvial plain, where energy to carry sediment eroded from the upper watershed is reduced. Here, the river tends to deposit the coarser sediment load produced and transported by the steep upper rivers, causing the river to fill and shift repeatedly. This filling and shifting results in a wide multiple-channel, dynamic system. As the coarser material drops out through this reach, the classic river system becomes a relatively stable reach with a

[3] Osterkamp constrained his regulated period to post-1926 because that is the year that American Falls Dam (RM 714) was completed upstream from Murphy, Idaho; this facility is capable of retaining 1,672,600 acre-feet of water (USBR 1998; Osterkamp 1997).

narrower, deeper river channel and meandering path. Most of the finer sediments and some of the coarser material passes through this zone, which has been called the transfer zone.

- As the river approaches the coastal zone (or lake, reservoir, or flat reaches along the longitudinal profile of the river), it often splits into a number of channels in the classic delta formation. As a river transports sediment, the velocity of flow decreases due to lower slopes and sediment is deposited. This process leads to the classic fan-shaped, multiple-channel delta configuration known as the deposition zone.

In this general classification, tributaries to the upper mainstream Snake River are located in the sediment production zone. Sediment production in the region tends to be controlled, in part, by large, infrequent storms that trigger mass failures and the subsequent delivery of sediment pulses from the hillslopes to the channels (Swanson et al. 1987). In addition to these storms, accelerated erosion from periodic wildfires, road building, and logging can also lead to dramatically increased hillslope sediment production (Swanson et al. 1987; Chapter 3), particularly in the mountainous areas of the basin.

In general, slope has a large impact on hillslope erosion because it affects the amount of kinetic energy that is available for downslope transport. Figure 4.4 shows a breakdown of hillslopes into gentler versus steeper slopes for the major subbasins. Table 4.1 provides these data by percent of surface area associated with each slope class. The breakdowns reflect the following categories:

- Greater than 40 degrees (resistant lithologic control)
- Ranging from 30 to 40 degrees (angle of repose—transport dominated by gravity)
- Ranging from 10 to 30 degrees (where fluvial processes dominate active source and transport areas)
- Less than 10 degrees (peneplanes with low energy surfaces)

In the Snake River Plain, sediment delivery from hillslopes to the channels is somewhat limited. Upstream from the confluence of the Owyhee and Snake Rivers (RM 393), hillslopes are relatively gentle (almost 60% of the surface area is less than 10 degrees). Therefore, hillslopes in this region probably contribute a relatively smaller volume of sediment to the tributaries or mainstem as compared to steeper reaches such as in Hells Canyon. Also, where glaciation has formed valleys with broad, flat floors (such as in the Little Lost River in the eastern Snake River Plain), debris slides and debris flows in tributary channels commonly do not reach the main channel (Swanson et al. 1987).

However, it is important to note that even on gentle slopes such as those found in the Snake River Plain itself, erosion and heavy runoff can occur on gentle slopes just through rainsplash (Summerfield 1991). This is particularly true for soil surfaces that lack vegetative cover, such as agricultural fields that have not yet been planted or have been harvested, or surface irrigated agricultural lands as discussed in Chapter 2.

Figure 4.5 shows the stream profile for the mainstem Snake River from its headwaters near Jackson Lake, Wyoming (RM 989) downstream to past Anatone, Washington (RM 145). The profile and nature of the Snake River Plain upstream from the Owyhee River confluence

(RM 393) suggest that this reach is located in the sediment transfer zone (Figure 4.3), where much of the coarse material that reaches the Snake River from the production zone drops out and most of the finer sediments passes through. Osterkamp (1997) argues that the lack of tributary inputs of sediment and water have resulted in “little geomorphic change in this part of the Snake River bottomland since the [Bonneville] flood” in this reach.

Again, because this portion of the study area is highly regulated, these projects have undoubtedly altered the pre-regulation delivery of water and sediment, both of which can affect the channel and valley morphology. For example, since “development and construction of dams and reservoirs on the Snake River and its major tributaries, the coarse sediment that was previously available for transport mostly has been stored in those reservoirs upstream from Swan Falls” (Osterkamp 1997), as discussed further in the following section.

However, in contrast to the next downstream reach, landforms in this reach have been relatively unaffected by the changes brought on by the current climate conditions and anthropogenic disturbances. In the reach upstream from the Owyhee River confluence (RM 393), islands within the channel formed as the slackwater deposits were incised and these relict landforms “have not been substantially altered in shape or location by recent (i.e., post-flood Holocene) geomorphic processes” (Osterkamp 1997). In contrast, the reach downstream of the Owyhee River to Farewell Bend has been “strongly affected by tributary inputs of water and sediment, and extensive evidence of post-Bonneville Flood erosion and deposition is apparent” (Osterkamp 1997).

4.1.1.3. Sediment Dynamics

Knighton (1998) summarized earlier research that defines three basic components for sediment loads:

- Dissolved load—material transported in solution
- Wash load—particles readily moving in suspension that are usually finer than material found in the bed
- Bed-material load—all sizes of material found in appreciable quantities in the bed

Within the bed-material load, bedload (rolling, salting, or sliding on the bed) is further distinguished from suspended load (turbulent mixing processes within the flow). From a geomorphologic standpoint, the bedload has a relatively greater influence on river channel form than suspended load.

As is common for most watersheds, the available sediment data are predominantly measurements of suspended sediment concentrations and loads, while bedload data are rare. Sediment yields and grain size information from available studies are summarized where applicable, and sediment concentrations and loads are presented for available datasets. Figure 4.6 provides a snapshot of the relative median particle size (d_{50}) in the channel bed throughout the study area.

In basins upstream from the Owyhee River (RM 393), sediment yields range between 3.9 and 61 tons per m^2 per year (tons/m^2/year) (or 0.005 to 0.071 cubic yards/acre/year) (USGS 1994, 1995). Osterkamp (1997) interprets these data to mean that "sediment concentrations in streamflow and sediment yields are low relative to many other drainage basins of similar size and climate." USGS (1994, 1995) data indicate that loads of suspended sediment are mostly in the silt-clay size fractions (<0.062 mm [0.002 inch]), and only a limited supply of sand (0.5 to 2 mm [0.02 to 0.07 inch]) is available for transport as bedload.

The dominance of fine particles in the suspended load is likely due, in part, to numerous water storage projects in this area, such as the American Falls Reservoir (RM 714). Clark (1994) states that "reservoirs on the mainstem of the [upper] Snake River probably trap...most of the suspended sediment load generated from upper parts of the basin." Below C.J. Strike Reservoir (RM 494), samples of bed and bank material support the USGS conclusion that little sand is available for downstream transport. Bed sample classifications (based on the subsurface sediment layer below the surface bed layer) indicate that gravels, cobbles, and boulders/bedrock make up more than 99% of bed material between RM 492 and RM 470 (Figure 4.6; IPC 2000b). This study also concluded that the surface bed materials are sufficiently well armored that the riverbed is expected to be stable under both C.J. Strike-regulated and hypothetical free-flowing conditions (IPC 2000b).

Data collected from bank samples between RM 493 and RM 465 downstream of C.J. Strike indicated that the majority of mainstem bank material was classified as sandy clay, low-plasticity silt, or poorly sorted sand to sandy clay (IPC 2000b). Although bank material may be available for transport to downstream reaches, relatively low TSS concentrations suggest that only minor volumes of sand-sized particles are eroding from the banks in the reach of the mainstem upstream from the Owyhee River (RM 393). For example, water quality data from Murphy (average TSS of 30 mg/L and average of 76% finer than 0.062 mm [0.002 inch]; RM 452; USGS 13172500) indicate that very little sand material is being transported downstream in the suspended load (USGS 2001b).

In addition to retaining a portion of the suspended sediment load, reservoirs also have reduced peak flows that typically transport most of the bedload. These flows control channel morphology because the largest sizes of transported sediment and the greatest rates of sediment discharge occur during annual flow peaks (Knighton 1998). Osterkamp (1997) argues that "the movement of 20- to 30-mm (0.78- to 1.81-inch) size gravel is limited [in the eastern Snake River Plain] … owing to a limited availability of these sizes for transport and because the reduction of peak flows by regulation reduces the transport capacity of those sediment sizes by streamflow." Osterkamp's analysis was based on hydraulic modeling performed by the USGS (Kjelstrom 1992). Upstream from the confluence of the Owyhee River, the bedload flux "may be low owing to a limited supply of sediment as well as to limited transport capacity." For example, bedload sediment transport is restricted by the low gradient in this reach. Osterkamp (1997) calculated the stream gradient in this reach to be 0.8 m (1.21 ft) per mile (0.0002), which is an extremely low slope that is almost half the gradient of the next downstream reach between the Owyhee and Weiser rivers (0.6 m [1.98 ft] per mile, or 0.0004).

4.1.2. Major Tributaries Between Owyhee River (RM 393) and Weiser River (RM 351)

Downstream of the Owyhee River and upstream from the HCC, the nature of the mainstem Snake River changes because of tributary inputs from the Owyhee, Boise, Malheur, Payette, and Weiser rivers. All of these tributaries are regulated to some degree and they have a significant influence on the mainstem immediately upstream from the HCC. Pre- and post-regulation conditions are evaluated in more detail in this section. (None of the post-regulation conditions are related to any of the HCC operations.)

4.1.2.1. Stream Flows

The collective drainage areas of the Owyhee, Boise, Malheur, Payette, and Weiser rivers comprise 36% of the 179,200-m^2 (69,200-mi^2) drainage area to Brownlee Reservoir as measured on the Snake River near Weiser, Idaho (Figure 1.1). They comprise 92% of the drainage area between Brownlee Reservoir (RM 335) and Swan Falls Dam (RM 458), which is the next upstream dam on the Snake River. Current average flows increase by 64% from 10,930 cfs at Murphy (RM 452) to 17,940 cfs near Weiser (RM 351); discharge inputs from each of the major tributaries are summarized in Table 4.2 and shown in Figure 4.7. Between the Owyhee and Weiser river confluences, the Owyhee, Boise, Malheur, Payette, and Weiser rivers supply approximately 83% of the total inflow, with the Payette and Boise rivers contributing most of the increase (Table 4.2).

Regulation projects on each of the tributaries have altered pre-regulation conditions; an example was previously shown for the mouth of the Boise River in Figure 2.3. In general, regulation projects have reduced peak flows, similar to the upper mainstem reach. The specific effects on post-regulation sediment transport are discussed later in this section.

4.1.2.2. Morphology

On a basin-wide scale, the Owyhee, Boise, Malheur, Payette, and Weiser rivers can be generally classified as sediment production zones. On a smaller spatial scale, Schumm's river classification can also apply to individual watersheds of these tributaries. For example, the Boise River originates in the steep central Idaho mountains, where a large volume of sediment is produced (IDEQ 2000). As the Boise River leaves these mountains, it enters a lower gradient reach on the Snake River Plain that causes sediment material to be deposited prior to reaching the mainstem Snake River. During the Pleistocene, this material was deposited into extensive gravel terraces that line the Boise River system (Othberg 1994).

Within the tributary sediment production zones, the Boise and Payette rivers have the highest proportions of steep slopes. Mass failure processes that are most common in these types of areas include rapid-failure debris slides and debris flows, and relatively slow-failure slumps, earth flows, and soil creep (Swanson et al. 1987). These mountainous basins are also areas that have been affected by fires, logging, and mining practices. Some studies have concluded that soil movement by debris slides in non-forested clear-cutting and broadcast burning areas may be two to four times higher than in forested areas (Swanson et al. 1987). Within the drier part of the study area (see Figure 3.4), dry gravel and rock slides accelerate transport on steep slopes (greater than 60% slope gradient) following wildfires or prescribed burning (Megahan and Molitor 1975 as referenced in Swanson et al. 1987). Swanson et al. (1987) also state that “slide erosion from [logging] roads can be several times higher than in forested areas.” The lack of a well-developed soil horizon in this area (see Section 3.4) may be further evidence that mass failure and downslope debris flow events are relatively common. Although some of the effects of fires are natural, the anthropogenic influences on burning rates and deforestation are discussed in more detail in Chapter 2.

Historically, sediment in these tributaries had the potential to be transferred through to the Snake River (the influence of the dams on sediment transport under current conditions is discussed in the following section). Drainage densities (the sum of all stream lengths divided by basin area) and relief ratios (the difference in elevation from highest to lowest elevation, divided by basin length) can be used as indicators of basin energy and potential transport efficiency. Relief ratios are analogous to the length-stream factor in the revised universal soil loss equation, where a higher erosion potential is associated with longer and steeper slopes (Wischmeier and Smith 1978). Figure 1.1 shows relative drainage density patterns and Table 4.3 provides relief ratios for each of the major tributaries.

These data show that the Idaho tributaries (Weiser, Payette, and Boise rivers) have relatively higher relief ratios (as compared to the Malheur and Owyhee rivers) that reflect a higher sediment transport potential. This trend can also be seen in the tributary stream profiles shown in Figure 4.8 and the hypsometric curves in Figure 4.9. These figures show that the higher elevation reaches have the most potential to transport sediment (i.e., higher elevation profiles are more concave) and that the tributaries from the Idaho side tend to have a larger percentage of their drainage area at higher elevations. In addition, the significant change in gradient between the headwater areas and the lower reaches on the Snake River Plain suggest that the lower reaches act as a sediment trap. For example, the gradient in the lower Boise River (0.002, or 4 m [13 ft] per mile) is significantly less steep than the headwater gradient in the upper Boise River (0.010, or 16 m [52 ft] per mile; Figure 4.8). This marked change suggests a reduction in sediment transport capacity, as discussed in the following section.

4.1.2.3. Sediment Dynamics

4.1.2.3.1. Pre-regulation Conditions

In contrast to the upper mainstem reach, in the mainstem reach between the Owyhee and Weiser rivers “movement of the 20- to 30-mm (0.78- to 1.81-inch) gravel sizes is inferred to be limited mostly by transport capacity and not by a limited supply of sediment” (Figure 4.6; Osterkamp 1997). That is, relative to the upper mainstem reach that does not have significant sediment inputs from its tributaries, the major tributaries in this reach have the potential to deliver much larger sediment loads.

To address the inputs from the tributaries, Osterkamp evaluated USGS data collected in the Boise and Payette rivers. He concluded that “60 to 70 tons per m^2 per year [0.069 to 0.081 cubic yards/acre/year] may approximate the long-term average of sediment yielded to the Snake River by streams from the mountainous areas of Idaho.” This is only slightly lower than historical measurements made between 1939 to 1940 upstream from Arrowrock Reservoir on the Boise

River near Twin Springs, Idaho, which indicate an annual sediment load of 99 tons/m^2/year (0.115 cubic yards/acre/year; USGS 1940). Osterkamp further estimated pre-regulated "bedload yields to the Snake River of about 1 ton/m^2/year [0.001 cubic yards/acre/year] of gravel (sizes of 2 to roughly 30 mm) and 10 tons/m^2/year [0.012 cubic yards/acre/year] of sand (sizes of about 0.3 to 2 mm)" (Osterkamp 1997). These estimates are consistent with a computed sediment load of 8.8 tons/m^2/year (0.010 cubic yards/acre/year) for medium- and coarse-sand sediments (greater than 0.5 mm [0.002 inch]) in the Boise River from 1911 to 1995 (Emmett pers. comm. as referenced in Osterkamp 1997).

These historic sediment yield estimates are consistent with recent water projects on the Payette River, which indicate that sediments continue to be produced in headwater areas. A hydroelectric facility was built on the Payette River near Horseshoe Bend in the early 1990s and the resulting change in gradient has caused sediment to accumulate in the canal upstream from the powerhouse. Between 1997 and 2000 nearly 388,550 cubic yards (241 acre-feet, or 60 acre-feet/ year [0.13 million tons/year]) of material have been dredged (Buchanon, pers. comm. 2001). Although studies are currently ongoing to estimate what percent of the total river load the dredged load accounts for, this example illustrates that high volumes of sediments continue to be produced in headwater areas upstream from regulation projects.

Suspended sediments appear to be the dominant component of sediment loads in the headwaters of the major tributaries. For example, based on USGS data collected in 1994 from the Middle Fork of the Boise River near Twin Springs, Idaho, (USGS 1318500) computed suspended load was 81% and bedload was 19% of the total sediment load (USGS 1994). These proportions are close to typical literature estimates of bedload comprising 10 to 15% of the total sediment load (Morris and Fan 1998), and only slightly higher than reported bedload proportions of 2 to 10% at Anatone, Washington (Jones and Sietz 1980). Suspended sediment transport is mostly associated with high-magnitude discharges (Knighton 1998). A local example of this is Emmett's (1975) analysis in the Upper Salmon River, which originates in the Idaho Batholith similar to the Boise and Payette rivers. He estimated that about 75% of the total suspended sediment load is transported during the 4% of the time at which flows are above bankfull (Emmett 1975).

4.1.2.3.2. Post-regulation Conditions

These numbers represent historical pre-regulated sediment conditions; however, water regulation projects on each of the major tributaries have markedly reduced available sediment that reaches the mainstem. Following regulation, reservoirs on each of the major tributaries (except the Weiser River) have retained sediment that can no longer be transported downstream into the Snake River mainstem and Hells Canyon. For example, Arrowrock and Lucky Peak reservoirs on the Boise River each trap between 85 to 90% of incoming sediment at average to full lake capacities on an annual basis (USBR 2001).

Within Regulation Projects

With respect to particle size, when a tributary enters an impounded reach, the bedload and coarse fraction of the suspended load generally are deposited immediately to form topset delta deposits. Although reservoir sediment transport dynamics rework this coarse sediment to some degree, the majority of sediments within bottomset beds near the dam face tend to be finer particles (Morris and Fan 1998). Therefore, it is reasonable to assume that the majority of sediments released past reservoirs, either during low flow years or as part of routine operations, are predominantly finer particles (silts and clays) from the center of the reservoir, where velocities and sediment transport capacities are highest.

Two examples from the Boise and Payette rivers support the assumption that the majority of particles retained (and subsequently released) within the reservoirs are finer materials. Based on USBR analyses of surface sediment within Arrowrock Reservoir, samples collected from the center of the reservoir were predominantly composed of silts and clays (an average of 95% was <0.062 mm [<0.002 inch] in diameter) (Figure 4.6; USBR 2001). Surface sediments collected from the Black Canyon Reservoir on the Payette River show a similar trend. Two samples in the upstream third of the reservoir were composed primarily of sand (an average of 32% was <0.50 mm [<0.02 inch] in diameter; USBR 1984). In contrast, four samples collected from the downstream third of the reservoir were composed primarily of silts and clays (an average of 62% was <0.074 mm [<0.003 inch] in diameter; USBR 1984).

Downstream from Regulation Projects

Downstream from the water regulation projects, additional fine- and coarse-grained material undoubtedly enters the tributary channels and joins the fine sediments released from the projects, in part, because the lower reaches are heavily farmed and grazed. By blocking off headwater areas, water regulation projects have also altered the parent rocks that produce sediment that can actually reach the mainstem. Based on the assumption that channel sediment composition likely reflects the surrounding lithologies, Table 4.4 summarizes the dominant lithologies upstream and downstream from storage for each tributary. Data in this table show that the dominant lithologies of source sediments downstream from storage (i.e., available for transport to the mainstem) differ markedly from the lithologies that would be available if storage projects were not in place. For example, in the Owyhee River watershed, although basalt and rhyolite dominate the basin as a whole, the 2% of the basin downstream from storage is made up primarily of sands associated with alluvial/eolian and sandstone rock types. In contrast, in the Boise and Payette watersheds the total basin is dominated by calc-alkaline intrusive materials (i.e., the granitic Idaho Batholith), while 31 and 18%, respectively, of the area downstream from storage is dominated by sands (69 to 79%, respectively) associated with the alluvial/eolian and sandstone lithologies. Although relative productive rates are unknown, on a semi-quantitative level based on relative drainage areas, storage likely has been effective at cutting off high-producing sediment supplies in the major tributaries.

In addition, available sediment in the lower reaches of these tributaries must travel through a gentle, sloping plain toward the mainstem of the Snake River. This extensive plain likely acts as a sediment trap because the gradient is so much lower in the lower reaches than the adjacent highlands. It is likely that most of the sediment that is deposited within the lower reaches consists of coarser sediments that cannot be transported across the low gradient. Therefore, only a minimal volume of coarse bedload (including sands and gravels) probably enters the tributaries.

An even lower proportion of the coarse materials reaches the mainstem Snake River. Suspended loads measured at the mouths of the tributaries generally confirms that the majority of suspended sediment that reaches the mainstem consists of finer-grained materials (silt-clays). For example, between 1974 and 2000, USGS (2001a) collected TSS data from the Boise River near Parma (USGS 13213000). Of 130 samples collected for TSS, 98 samples were also sieved through a 0.062-mm (0.002-inch) mesh to determine the percentage of silt-clay. These data indicate that silt-clay make up an average of 81% of suspended sediment delivered to the mainstem.[4] Therefore, only 19% of the suspended sediment load discharging into the mainstem is composed of sands. This proportion would probably be only slightly larger if bedload data were available, as bedload may only comprise up to 20% of the total sediment load (USGS 1994).

Table 4.5 presents flow data and TSS concentrations for the mouths of the other major tributaries and the mainstem Snake River in this reach (bedload data are not available). These data were collected by USGS (1997a and b) and by IPC for various periods of record (POR). Given that these five tributaries drain over 64,750 km^2 (25,000 mi^2), the suspended sediment loads seem to be relatively small. Also, post-regulation average annual loads for the Boise and Payette rivers (approximately 30 tons/m^2, or 0.035 cubic yards/acre) are less than half of the pre-regulation loads estimated by Osterkamp (1997) and measured by USGS (1940). The transport and delivery of these sediments in the mainstem Snake River are discussed further in the following section.

4.1.3. Mainstem Between Owyhee River (RM 393) and Brownlee Reservoir (RM 335)

This reach of the mainstem Snake River directly affects the delivery of water and sediment into the HCC. The portion of the mainstem between the Owyhee and Weiser rivers contains the majority of channel islands associated with the Deer Flat National Wildlife Refuge (DFNWR). The DFNWR was studied extensively by Osterkamp (1997) as part of the Snake River Basin Adjudication process. Between Weiser (RM 351) and Brownlee Reservoir (RM 335), less-detailed information is available but it is summarized as appropriate. Although the headwaters of Brownlee Reservoir vary depending on drawdown conditions, RM 335 is generally considered to be the beginning of the riverine section of the reservoir (IPC 2000a).

4.1.3.1. Stream Flows

Under post-regulation conditions, average flows increase by 64% from 10,930 cfs at Murphy (RM 452) to 17,940 cfs near Weiser (RM 351), with the majority of inflow provided by the major tributaries discussed in the previous section (Table 4.2). In the 16 river miles between the confluence of the Weiser River (RM 351) and the headwaters of Brownlee Reservoir (RM 335), flows do not significantly change. Thus, flows at Weiser are generally used to estimate inflows into the HCC. After flowing past Weiser, the Snake River brings an average of almost 13 million acre-feet annually into the HCC (Figure 4.2; USGS 2001b).

A comparison of observed and estimated unregulated inflows to Brownlee Reservoir is provided in Section 4.2.

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[4] In the draft report, it was reported that silt-clay particles made up an average of 79% of suspended sediment based on data through 1997 (USGS 1997a). The 81% value used in this report includes suspended sediment data collected by the USGS through 2000.

4.1.3.2. Morphology

Within Schumm's classification, this reach of the mainstem can be classified as a sediment transfer zone (Figure 4.3), similar to the upper mainstem reach. However, relative to the upper mainstem reach, the Snake River in this reach has been more "strongly affected by tributary inputs of water and sediment [from the Owyhee, Boise, Malheur, Payette, and Weiser rivers] and extensive evidence of fluvial erosion and deposition is apparent [since the Bonneville Flood]" (Osterkamp 1997). These processes are consistent with a transitional zone where the channel adjusts to new gradients and confinement conditions.

Upstream from the HCC, the riverbed and banks interact dynamically within a relatively self-formed alluvial plain. Near Weiser, the broad river valley (about 185 miles wide) has an extensive floodplain developed for agriculture. The valley and river channels are relatively wide, averaging about 3,200 m (10,500 ft) (3 km [2 mi]) and 300 m (1,000 ft) from right to left bank, respectively. The longitudinal slope of the river gradient is very shallow, averaging only about 0.6 m (1.98 ft) per mile (0.0004) near Weiser (Figure 4.5).

Tributary and mainstem dams have reduced peak flows and decreased the volume of sediment transported downstream past Weiser. Channel islands of the Snake River between Swan Falls and Farewell Bend have increased in areal extent by an average of 8% since 1938 (Johnson and Dixon 1997). This aggradation is "inferred to be a reaction to the conditions of regulated streamflow" (Osterkamp 1997). Johnson and Dixon (1997) concluded that "the increasing size and headward expansion of the islands suggest that either flow during the past half of the century has become insufficient to move sediment from shallow areas of the channel and/or that sediment supply has increased such that it exceeds the power of the stream to transport it."

It is important to note that the aggradation in the reach upstream from the HCC is believed to be a result of both upstream regulation projects and the relatively low gradient. Aggradation is not believed to be occurring because of backwater effects from Brownlee Reservoir, the upper end of which is located more than 80 km (50 mi) downstream from the Owyhee River confluence. This is discussed in more detail in the following section.

4.1.3.3. Sediment Dynamics

4.1.3.3.1. Within DFNWR

Some portion of sediment that enters this reach of the mainstem both from the upper mainstem and from the major tributaries appears to be deposited around numerous channel islands within the DFNWR. Channel islands in the DFNWR reach of the Snake River have increased in areal extent by an average of 8% since 1938 (Johnson and Dixon 1997) as a result of "adjustment to the altered discharges and flow durations of the last 70 years" (Osterkamp 1997). The DFNWR islands are comprised of various sediment sizes; including Bonneville Flood slackwater deposits (d_{50} between 0.08–0.38 mm [0.0031–0.015 inch]), modern gravel deposits (d_{50} of 3.0 to 40 mm [0.12 to 1.6 inches]), and Bonneville Flood gravel deposits (d_{50} of 40 to 79 mm [1.6 to 3.1 inches]; Figure 4.6; Osterkamp 1997).

To determine what sizes of particles might be transported around these islands under current post-regulation conditions, hydraulic modeling was conducted for representative islands (Blind, Cigar, Silo, Suzy, and Ketchup Islands) (Osterkamp 1997, Kjelstrom 1992). Modeling results indicate that, at typical peak flow discharges, nearly all of the 0.1-mm (0.004-inch) fine sand likely is transported as suspended load and the 1.0-mm (0.04-inch) medium sand likely is transported entirely as bedload (Osterkamp 1997). As expected, during low flows downstream transport of the mobilized sand is limited because "numerous secondary channels… as well as gravel bars … have water that is too shallow to provide water velocity great enough to transport the sand sizes, and thus, deposition is likely to occur" (Osterkamp 1997).

Within the mainstem channel, measurements and observations by Osterkamp (1997) indicate that median bed particle sizes (d_{50}) are about 6 mm (0.24 inch) where material is being moved either by historical pre-regulation or modern post-regulation transport processes (in contrast, median bed material sizes of Bonneville Flood debris that has remained in place range upwards of 100 mm [3.9 inch] or more). Within the channel, Osterkamp (1997) observed that sands and fine gravels that currently move under non-flood conditions have d_{50} values that range between 0.23 mm and 3 mm (0.009 and 0.12 inch), while sands and fine gravels that move under peak flows are generally no larger than 30 mm (1.2 inches) (Figure 4.6).

Osterkamp (1997) also states that movement of material in the 20-30 mm (0.79–1.2 inches) range is restricted by limited transport capacity resulting from the other upstream regulation projects. Specifically, the USGS modeling results predict that 20-mm (0.79-inch) gravel would only move under peak flows (in the 25–35,000 cfs range) in isolated pockets as specific stream hydraulic conditions allow. Larger gravels (30 mm [1.2 inches]) would only be mobilized at high peak flow conditions (35,000 cfs; Osterkamp 1997).

Based on observation by IPC, it appears that although this material might be able to be mobilized if it were available, in fact the channel bed in this reach is armored in many places by baseball- and softball-sized materials. For example, Photo 5 shows the condition in 1999 of mainstem gravel bars and mid-channel bars at RM 357.1.

Photo 5. Surface and subsurface bed materials collected at RM 357.1.

Subsurface materials below surface gravels were submitted for particle size distribution (laboratory reports are contained in Appendix D of Parkinson et al. [2003]). These results indicate that the d_{50} for these materials ranges between about 5 mm (0.20 inch) (RM 380.1) to 25 mm (0.98 inch) (RM 368.0). Although the surface gravels were not submitted for analysis and no pebble counts were performed, the photographs above suggest that surface gravels are indeed baseball and softball-sized materials that likely armor the smaller subsurface materials.

Gravels are being deposited on channel islands in this reach based on Osterkamp's (1997) measurements, as well as IPC observations (e.g., a newly-formed gravel bar at RM 380.1). Thus, under post-regulation conditions within this reach, most of the gravels appear to be re-deposited a relatively short distance downstream from where they were entrained. These observations and measurements suggest that high volumes of gravels in this reach do not travel downstream past the Weiser gage because post-regulation flows are rarely high enough to mobilize the surface gravels. In addition, the low mobility of these gravels suggests while sand materials within the surface layer are available for bedload transport, significant volumes of subsurface sand materials are probably not transported downstream.

4.1.3.3.2. Near Weiser

The potential movement of sands and gravels from the DFNWR downstream past Weiser was evaluated using available TSS data and channel bed observations. Downstream of the Weiser River confluence (RM 351), an average of 1.09 million tons/year (503 acre-feet/year) of suspended material are transported. Analysis of the USGS data through 2001 indicates that, on average, about 81% of the suspended sediment in the Snake River near Weiser is smaller than 0.062 mm (0.002 inch), meaning that this sediment is predominantly silts and clays (smaller than very fine sand). This high proportion of silt-clay suspended sediment is likely a result of both upstream regulation that retains coarser particles and agricultural and grazing land use inputs.

Although the USGS dataset at Weiser does not include bedload sediment measurements, if sand and gravel is being transported beyond the islands in the DFNWR, then changes in channel morphology would be expected downstream near Weiser. More specifically, high bedload transport typically would be reflected in long-term changes in cross sectional area and stage-discharge relationships, as well as backwater effects associated with Brownlee Reservoir. None of these typical outcomes appear to be occurring; thus, significant volumes of sand and gravel are likely not being transported out of the DFNWR reach and into the HCC. Each of these items is discussed in more detail below.

USGS cross sections surveyed at Weiser indicate that there has been no significant change in channel cross sectional area for more than 60 years (Parkinson et al. 2003). As expected, the thalweg has migrated laterally, but there does not appear to be any major aggradation of the channel at this location. This suggests that the mainstem has not been carrying a significant load of coarse material past Weiser since at least the 1940s. In addition, an evaluation of the stage-discharge relationship would be expected if a high bedload was causing the channel to change form. The seven rating curves that have been developed by the USGS since 1910 essentially plot on top of one another (Parkinson et al. 2003). This suggests that the river near Weiser has neither degraded nor aggraded substantially since the gage was installed 90 years ago. Note that even when a riverbed is immobile and the bed materials is not available, sediment can still be transported through that reach. The stability is only used to illustrate the fact that there is no sign of significant aggradation or degradation of the river at this locations.

4.1.3.3.3. Downstream from Weiser

If a significant volume of sands and gravels was being transported past Weiser, then these materials would be deposited either within the Brownlee Reservoir delta or within the reservoir backwater area. Typically, backwater effects associated with an increase in the upstream water surface behind a reservoir cause the velocity of the incoming river to decrease as it enters the reservoir. This resulting decrease in velocity causes the deposition of relatively coarser materials within the reservoir delta.

As discussed in more detail in Section 4.2., deep core samples collected in the reservoir delta between RM 324 and RM 320 consisted almost entirely (99%) of very-fine sand (0.125–0.063 mm [0.005–0.002 inch]) and silt-clay (≤0.063 mm [0.002 inch]; Parkinson et al. 2003). Shallow surface samples collected upstream from the delta at RM 335 and RM 330 consisted primarily of silt-clay with some sand and gravel (Parkinson et al. 2003). The trace volumes of coarser materials were only found after numerous deployments to seek out pockets of coarse sediment along existing channel margins; the majority of deployments in this reach hit bedrock or armored bed surfaces that produced no recoverable sediment (Parkinson et al. 2003). This study suggests that sand and gravel are not being deposited in areas associated with the Brownlee Reservoir delta.

In addition to the reservoir delta information, the potential deposition of sands and gravels in the backwater area was also evaluated. As the reservoir delta forms, the areas where these coarse materials are deposited tend to migrate upstream as the delta aggrades (Morris and Fan 1998). A local example is the Black Canyon project on the Payette River. Between 1924 and 1971, the average rate of sedimentation was 336 acre-feet/year (0.73 million tons/year), the majority of which was deposited at the upper end of the reservoir (USBR 1984). The build-up of sediment effectively filled in the area where the streambed entered Black Canyon Reservoir. The resulting flooding prompted about 4.5 km^2 (1,100 acres) of land upstream from the reservoir in the Montour Valley to be acquired by federal agencies to be managed as a wildlife and recreation area (USBR 1994).

In the Snake River, a similar backwater effect would be expected if the mainstem was carrying a significant bedload of coarse material in this reach. Specifically, sand and gravel deposition and bar expansion would be reflected in photographs and associated riparian vegetation changes. To support a wildlife habitat assessment, oblique photographs from the early 1900s and 1999 taken between Weiser (RM 351) and Farewell Bend (RM 335) were compared (Blair et al. 2001). These photos indicate that the river channel substrate and morphology did not change significantly for both periods at Weiser, Westlake Ferry Crossing (RM 345), and Farewell Bend (RM 335; Blair et al. 2001). For example, at the Westlake Island Ferry Crossing the channel was characterized as wide and lacking small channel islands in both 1908 and 1999. A comparison of aerial photographs taken in 1955 and 1999 at all three locations resulted in similar conclusions. Although the floodplain characteristics changed because of agricultural land uses, the channel morphology was comparable for both periods (Blair et al. 2001). An example from the Westlake Island Ferry Crossing is shown in Photo 6 (Blair et al. 2001).

These photographs, in conjunction with the apparent stability of the Weiser cross section and stage-discharge relationship, suggest that the reservoir backwater area has not migrated upstream to Weiser in the 40 years since Brownlee was completed. These results also suggest that substantial volumes of sand and gravel are not being transported past Weiser and into the reservoir backwater area. This is not to say that the HCC has not caused any sand and gravel to be trapped upstream. However, the proportion of coarse materials that may be deposited in the 26-km (16-mile) reach between the Brownlee Reservoir and Weiser is likely too small to feed the sediment requirements of the 61-km (100-mile) reach of Hells Canyon below the HCC (the potential contribution of coarse sediment that is trapped within the reservoir reach is discussed further in the following section). This conclusion is supported by additional analyses indicating that channel sediment in Hells Canyon continues to be derived predominantly by local canyon supplies, as discussed in Chapter 5.

Photo 6: August 23, 1955 (left) and May 7, 1997 (right) aerial photographs of Westlake Island (RM 345). Channel morphology appears to be similar in both views.

Another conclusion that can be drawn is that the construction of the HCC has not contributed to the aggradation seen in the DFNWR reach. When the Snake River portion of the DFNWR was created in 1937, 36 islands were designated for protection because they provided a riparian corridor in the desert (USFW 2001). Thus, these islands were present long before Brownlee Reservoir was completed in 1957. These observations strongly suggest that aggradation in the DFNWR is occurring primarily because of upstream regulation projects and the relatively low gradient in this reach (approximately 0.6 m [1.98 ft] per mile, or 0.0004), not due to the presence of Brownlee Reservoir.

Finally, the overall transport capacity of the Weiser reach was evaluated in the context of the downstream materials found in Hells Canyon. That is, in order for the Snake River upstream of the HCC to have been a significant contributor of coarse bed materials below the HCC, these sediments would have had to have been transported past Weiser above the HCC. As indicated previously, the bedload in this reach during the last 70 years or more has been restricted to sand and gravel sizes with a d_{50} of 30 mm (1.80 inches) or smaller (Osterkamp 1997). In contrast, the average d_{50} of the armored surface bed layer in Hells Canyon is much larger than 30 mm (1.80 inches) (Chapter 5). Because an armor layer can be developed when a river is incapable of moving the largest sizes of the bed material, a smaller size (for example, the d_{50} rather than the d_{90}) than the largest size of the armor layer was chosen to calculate whether Hells Canyon bed materials could have been transported through this reach under current hydrologic conditions.

Using data from underwater photography, IPC determined that the average d_{50} of the surface layer in Hells Canyon is 144 mm (5.7 inches) (Figure 4.6; Chapter 5). To move material of this size through the reach of the Snake River between the Weiser River and Brownlee Reservoir, the Snake River would have to flow at a depth of more than 60 m (200 ft). A rough estimate based on cross section and slope data indicates that the flows required to produce depths in this range are on the order of at least 9 million cfs (Parkinson et al. 2003). Alternatively, for a flow of about 100,000 cfs (in the range of the current 100-year peak flow), the slope would have to be on the order of 0.007 (11 m [36 ft] per mile) to mobilize sediments with a d_{50} of 144 mm (5.7 inches). This is almost 6 times steeper than either the current or pre-HCC gradient in reach extending from Weiser to Brownlee Dam (approximately 1.89 m (6.2 ft) per mile; Figure 4.5; USGS 1925).

Thus, it appears that the transport capacity of the river above the HCC is not sufficient to mobilize and transport material like that found in the riverbed of the Hells Canyon reach. Together, the weight of evidence indicates that the Snake River above the HCC is not capable (under either geologically recent or post-regulation hydrological conditions) of transporting the sizes of bed materials that are dominant in the river below the HCC. IPC recognizes that bedload in this reach may contribute to downstream suspended sediment loads. Although it is possible that the d_{50} observed in Hells Canyon (144 mm [5.7 inches]) reflects a winnowed historical supply from Weiser (i.e., the HCC has retained enough fines such that the downstream d_{50} has increased), additional analyses presented in Chapter 5 indicate that the channel sediment in Hells Canyon has been derived predominantly by local canyon supplies for a long (geologic timescale) period of time.

4.2. Within the HCC

4.2.1. Stream Flows

After flowing past Weiser, the Snake River brings almost 13 million acre-feet annually into the HCC (USGS 2001b). Prior to the completion of Brownlee Reservoir in 1958, the natural hydrograph had already been altered by more than 10 major upstream facilities on the mainstem, as well as by 35 major water storage projects on the tributaries (see Chapter 2). Reservoirs upstream of the HCC are primarily intended to provide irrigation water for numerous agricultural users along the Snake River Plain. This regulation alters natural flow conditions by storing incoming water and releasing water at regulated rates (see Figure 2.3 for a comparison of pre- and post-regulation hydrographs upstream from the HCC).

Figure 4.10 compares the observed annual average input and the estimated unregulated input of water to Brownlee Reservoir for water years 1993 through 2000. Observed flows were determined by IPC based on flows below HCD (USGS 13290450) and changes in headwater elevations in Brownlee, Oxbow, and Hells Canyon reservoirs. Unregulated flows were predicted by USACE using a hydrologic model based on data from upstream regulation projects (O'Brian, pers. comm. 2001).

Of the years depicted, the minimum annual volume occurred in water year 1994 (dry year) and the maximum volume occurred in 1997 (wet year)[5]. As illustrated by the plot, for the dry period from 1993 to 1995, almost half (46 to 49%) of the estimated naturally occurring annual volume of the Snake River was diverted for agricultural purposes. During the wet period between 1996 and 1998, approximately one-third (32 to 34%) of the estimated naturally occurring annual volume was diverted.

In addition to diverting naturally occurring flows, reservoirs upstream from the HCC have also reduced peak flow rates by retaining the water for later release to meet water supply needs. Figures 4.11 and 4.12 compare regulated and unregulated (estimated naturally occurring) peak flow rates into Brownlee Reservoir for dry and wet water years (1994 and 1997). As expected, reduction of the peak flow rate is more significant during periods of limited water. For dry year 1994, regulation of the system results in a peak flow rate that is 36% of the estimated naturally occurring peak flow rate. In 1997, the regulated peak flow rate is 54% of the estimated naturally occurring flow peak rate.

Finally, it is important to note that the maximum unregulated peak flow rate was estimated to be approximately 150,000 cfs in wet year 1997, which is well above any flow of record in the Hells Canyon reach of the Snake River (98,000 cfs recorded at the gage station below HCD; USGS 2001b). Estimates by Parkinson et al. (2003) indicate that at flow rates of 300,000 cfs, even very coarse gravel (greater than 50 mm [1.9 inches]) in spawning beds is generally not mobilized within Hells Canyon; this is discussed in more detail in Chapter 5. This suggests that estimated unregulated peak flows would not be large enough to mobilize spawning gravels in the canyon.

Currently, the water level in Brownlee Reservoir fluctuates and the reservoir responds to the varying inflow and reservoir operation rules that constrain hydroelectric production, flood control, fisheries, and other uses (Parkinson et al. 2003). In terms of reservoir storage, total active storage capacity of the HCC (Brownlee, Oxbow, and Hells Canyon reservoirs) is 1.48 million acre-feet. This means that the HCC is only able to hold approximately 11% of the average annual inflow to the reservoir, which is markedly lower than other systems of similar size (Chapter 2). In addition, the total reservoir storage capacity in the Snake River Basin upstream from the HCC exceeds 10 million acre-feet. Thus, from a storage capacity perspective, the HCC represents only about 13% of the total storage capacity in the Snake River Basin.

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[5] Dry or wet designation of these water years is consistent with the hydrologic and hydraulic modeling report (Parkinson 2001).

Within the HCC reach, four major tributaries contribute additional inflow to the reservoirs. These include the Powder and Burnt rivers and Pine Creek from the Oregon side, and Wildhorse River from the Idaho side. Annual mean discharges for these tributaries, as well as daily minimum and maximum discharges, are shown in Figure 4.13. Numerous small or local tributaries also provide inflows, particularly during the spring snowmelt runoff period. All of these tributaries combined are estimated to contribute less than 1% of the total inflow into this reach (IDEQ 2001).

4.2.2. Morphology

In Schumm's general classification, the Snake River in the reservoir reach is located in the sediment transfer zone (Figure 4.3). Local tributaries—which include the relatively large Burnt and Powder rivers and the relatively small Wild Horse River and Pine Creek drainages—are steep and periodically erode and produce sediment. Coarse materials from local tributaries were typically deposited in this reach, as evidenced by islands and rapids that were present prior to the completion of the HCC (Blair et al. 2001). Some fine material (primarily sand) was also deposited, primarily in isolated areas where river hydraulic conditions allow (e.g., near-shore eddy environments). The HCC has altered these depositional patterns, as discussed in the following sections.

4.2.3. Sediment Dynamics

The sediments in the HCC are composed both of upstream mainstem sources and local tributary sources. Prior to the completion of the HCC, elevated loads of fine materials from anthropogenic disturbances may have washed through this reach into Hells Canyon. As discussed in the previous section, the majority of coarse sediment from upstream sources was either retained behind numerous other regulation projects or was deposited within the DFNWR reach. Coarse sediment from local tributaries appeared to be deposited in this reach in island and rapid channel features (USGS 1925; Blair et al. 2001). Following construction of the HCC, sediments that would have passed through to Hells Canyon have been retained within the three reservoirs. Most of the retained sediment is believed to be in Brownlee Reservoir because it is the uppermost and largest reservoir of the three.

4.2.3.1. Brownlee Reservoir

IPC recently undertook an extensive study to characterize the physical and mineralogical properties of sediment collected from the Brownlee Reservoir delta and thalweg (Parkinson et al. 2003). Results pertaining to sediment deposition within the reservoir are summarized in this section and the report in its entirety is included as an appendix to Parkinson et al. (2003). Three deep cores (up to 8.5 m [28 ft] in thickness) were collected from the reservoir delta between RM 324 and RM 320. The Snake River reservoir delta was targeted for sampling for two reasons: [1) because the delta environment is where flow velocities decrease and incoming bed load and coarse fraction of the suspended load are deposited (Morris and Fan, 1998); and 2) the Snake River is the largest tributary to the reservoir and was therefore considered the single greatest potential source of sediment.] In addition, shallow samples (up to 1.5 m [5 ft] in thickness) were collected every 8 km (5 mi) along the reservoir thalweg between RM 340 and

RM 285 near the dam. All samples were collected during low-flow conditions in the fall or winter while conditions favored deposition of fine sediments.

The deep delta cores were collected in areas where the maximum extent of coarse material was expected based on reservoir operations (Parkinson et al. 2003). Within the deep cores in the delta, very-fine sand (0.125–0.063 mm [0.005–0.002 inch]) and silt-clay (≤0.063 mm [≤0.002 inch]) made up 99% of the grain sizes (Figure 4.6; Parkinson et al. 2003). The remaining approximately 1% of the grain sizes included an overall broad suite of sand-sized grains decreasing in quantity (by weight) from fine sand through coarse- to very-coarse sand. These silt-clay sized materials had a specific gravity similar to that of a loamy, silty-clay, surface soil—averaging 2.54 and ranging from 2.36 to 2.74 (Parkinson et al. 2003). Bulk density in the deep cores ranged from 0.70 to 1.73 with an average of 1.32 grams per cubic centimeter (g/cm^3). The estimated bulk density averages were almost exactly the same as the typical bulk density (1.3 g/cm^3) as re-deposited sediments from cropland erosion (Trimble and Crosson 2000).

Using the 1980 eruption of Mount St. Helens ash as a depositional marker in the deep cores, sediment has accumulated at an average rate of about 0.3 m (1 ft) per year between 1980 and 1998 (Parkinson et al. 2003). The depositional rate within the reservoir likely has not been a continuous, steady-state phenomenon because of the variability of flows and sediment supply, fluctuating water-surface elevations, and changes in the reservoir-bed profile.

In addition to the deep cores, shallow surface samples were collected along the thalweg; these surface samples were divided into an upper-reservoir group (RM 335 to RM 312) and a lower-reservoir group (RM 310 to RM 285). In the upper-reservoir segment very little sediment was present between RM 335 and RM 330, where numerous deployments were required to recover trace amounts of sediment. The thalweg in this area appears to be composed either of hard boulder or bedrock surfaces or of consolidated cobbles and boulders (Parkinson et al. 2003).

Only limited volumes of sediment were able to be collected from specific target areas (e.g., depositional pockets near mid-channel islands). The grain size data showed a bimodal sediment distribution between RM 335 and RM 330, where the dominant silt-clay was mixed with small volumes of very coarse gravel (Parkinson et al. 2003). The bimodal condition indicates that the riverine portion of the upper-reservoir responds to a relatively stable flow regime (which periodically fluctuates between riverine and lacustrine conditions) punctuated by periodic higher flow velocity events (Parkinson et al. 2003). In contrast, surface samples collected between RM 325 and RM 312 contained no gravels and were dominated by silt-clay (average of 77%) and sand (23%).

Shallow surface samples collected from the lower-reservoir segment (RM 310 to RM 285) also are dominated by a silt-clay grain size that averages 99% by weight (Parkinson et al. 2003). Of the remaining 1%, the average percent of very fine sand is 0.5%, fine sand is 0.3%, and medium sand is only 0.1%. It is unclear what proportion of the sand is contributed by upstream sources versus local tributary sources. These data indicate that the lower part of the reservoir is under a relatively static lacustrine environment.

4.2.3.2. Oxbow and Hells Canyon Reservoirs

To estimate the long-term sediment deposition into the lower two reservoirs, recent bathymetry data were compared against pre-impoundment data (Parkinson et al. 2003). In the case of Wildhorse River, contour maps of the areas inundated by the reservoirs prior to filling were compared to bathymetry data collected in 1999 in the upper 7.2 km (4.5 mi) of Oxbow Reservoir. Initial data appeared to indicate a sedimentation yield from the watershed of about 0.22 acre-feet/mi^2 per year (479 tons/mi^2 per year; Parkinson et al. 2003). However, this preliminary estimate ignored the fact that the early contour maps stopped at the river's edge, while the bathymetry data continued through the full width of the reservoir.

During attempts to refine the estimate and account for the blank area in the early contour mapping, it became obvious that small adjustments in overlaying the recent bathymetry and the early contour data could easily overwhelm the differences between the two surfaces. This is partly a result of lack of detail in the early contour map (that is, the pre-impoundment 6-m [20-ft] contours are not sufficiently detailed on the valley floor, where most deposition would likely occur in this case). In addition, there are no common control points to allow accurate referencing of the two data sets. Similar constraints preclude a comparable analysis in Hells Canyon Reservoir. As a result, these data cannot be used to calibrate or verify sediment load estimates from Wildhorse River or Pine Creek.

4.2.3.3. Estimate of Sediment Yield from Local Unregulated Tributaries

Sediment supply from local tributaries into the HCC is another component of sediment production during extreme events in this reach. The Burnt and Powder rivers flow directly into Brownlee Reservoir, but both of these tributaries have dams located on them and thus some portion of the total sediment supply is blocked above the HCC. For example, out of a total drainage area of 2,850 km^2 (1,100 mi^2), 29% of the Burnt River watershed has been cut off from the lower reaches since 1938 because of Unity Reservoir (see Table 2.1).

These sediment yields are based on potential transport capacity[6]; for which a detailed methodology is described in the Sediment Report (Parkinson et al. 2003). Potential sediment transport was estimated only for flows occurring 1% or more of the time (i.e., a 1% recurrence interval or higher; Parkinson et al. 2003). Also, only those sediments larger than silt-clay (>0.063 mm [>0.002 inch]) are included. Again, it is important to note that these estimates cannot be confirmed or calibrated based on the earlier discussion of bathymetry issues of the lower reservoirs.

Wildhorse River and Pine Creek are the two largest tributaries entering the HCC that do not have large dams. Oxbow Reservoir has a very limited number of drainages that appear to supply sediment. The only other tributary that drains into Oxbow Reservoir (Salt Creek) that fit within the parameters of the methodology (i.e., basin size, percent slope, etc.) shows some evidence of sediment movement, but nothing like some of the tributaries to Hells Canyon Reservoir, the Snake River in Hells Canyon, or even tributaries to Brownlee Reservoir (Parkinson et al. 2003). These other tributaries show large amounts of obviously recently deposited materials of a wide

[6] A similar analysis conducted in the Hells Canyon reach downstream of the HCC is summarized in Chapter 5.

range of sizes from sands through cobbles with little vegetation growing on them. In contrast, sediment deposits at the mouth of Salt Creek are more armored and have large trees still growing over most of the alluvial fan area.

Wildhorse River (drainage area of about 458 km^2 [177 mi^2]) enters Oxbow Reservoir about 1.6 km (1 mi) downstream of Brownlee Dam. Sediment yield estimates for Wildhorse River indicate that no sediment would currently be supplied because the armor layer would not be broken up until flows reached about 1,270 cfs (Parkinson et al. 2003). Flows of 1,270 cfs are higher than the highest average daily flow with a 1% recurrence interval, which is about 1,340 cfs. However, the January 1997 flow in Wildhorse River was estimated to be about 4,200 cfs, which is well above armor mobilization. The movement of sediment at these high flows does not contradict the validity of the sediment transport equations because potential transport capacity was estimated only for flows occurring 1% or more of the time.

Pine Creek (drainage area of about 779 km^2 [301 mi^2]) enters Hells Canyon Reservoir just below Oxbow Dam. For events with a 1% recurrence interval or higher, sediment supply estimates for Pine Creek indicate that no sediment would currently be supplied because the armor layer would not be broken up until flows reached about 5,400 cfs (Parkinson et al. 2003). Flows of 5,400 cfs are more than twice as high as the highest average daily flow with a 1% recurrence interval, which is about 2,000 cfs. Similar to Wildhorse River, during the high flows in January 1997, Pine Creek moved a large volume of material, which would seem to contradict this conclusion. However, the January 1997 flow event had an estimated peak of 11,600 cfs, which is more than 5 times the 2,000 cfs flow with a 1.0% recurrence interval. At flow levels of 11,600 cfs, all sizes of sediments would be transported in Pine Creek (Parkinson et al. 2003). This appears to confirm, rather than contradict, the conclusion that negligible sediment supply appears to be currently available from Pine Creek for events with less than a 100-year return period.

4.2.3.4. Effects of the HCC on Downstream Sediment Supplies

The sediment budget presented in Parkinson et al. 2003 provides a conceptual framework that helps in identifying the potential effects of the HCC on downstream sediment supplies. As part of this budget, an estimate of between 1.47 and 2.78 million/tons per year of sediments would have washed downstream into Hells Canyon if the HCC were not in place. This estimate equates to between 279,000 and 387,000 tons per year of sand sizes or larger. Thus, the majority of this material (silt-clay) would likely have flushed through the canyon reach and does not represent the typical sizes of sediments that comprise locally important features such as sandbars and spawning sites (discussed in more detail in Chapter 5).

In addition, sediments are produced by the largest unregulated tributaries and retained in the reservoirs (11.4 million tons/year; Parkinson et al. 2003). This can be compared against the estimated downstream coarse sediment supply (8.6 million tons/year; Parkinson et al. 2003), as discussed in more detail in Chapter 5. If additional coarse sediments in the original channel bed in the reservoir reach were similar to bed materials observed upstream from the reservoir, these sediments were probably not very mobile. Only a portion of these coarser materials likely would have eventually passed through the 141-km (88-mi) reservoir reach to Hells Canyon.

Finally, Brownlee Reservoir has only trapped 8% of the total volume of sediment accumulated within the basin upstream from the HCC. Thus, the fine and coarse sediments that have been retained upstream as a result of the HCC are only a small fraction of the pre-regulation sediment supplies in the basin that might have otherwise reached Hells Canyon.

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5. HYDROLOGY, MORPHOLOGY, AND SEDIMENT DYNAMICS DOWNSTREAM OF HELLS CANYON DAM

Chapter Summary

Although stream flows and sediment dynamics influence geomorphology, the channel in Hells Canyon is controlled, in large part, by the imprint of its complex geologic history. In terms of stream flow (Section 5.1.), the HCC augments downstream flows by season depending on flood control, activity of resident fish, power requirements, and recreational constraints. Current peak discharges are not markedly different from pre-HCC flows because the HCC is capable of storing only approximately 11% of the river's average annual flow. Therefore, the HCC has relatively little ability to regulate floods because the reservoirs fill so rapidly.

Downstream of the HCC the river morphology (Section 5.2.) is characterized by a deep, steep, and narrow valley that is confined by bedrock walls, talus slopes, debris flows, landslides, and alluvial terraces (Bonneville Flood deposits, landslide backwater deposits, relict alluvial fans, and relict bars). This confinement has precluded the development of an alluvial floodplain that is typical of other rivers of comparable discharge area and gradient. Rapids and pools caused either by boulder bed material or by debris-flow fans are common in Hells Canyon.

Local tributaries and adjacent hillslopes together are estimated to produce a minimum of 8.6 million tons of sediment on an annual basis (Section 5.3.). These sediments are supplied directly from the tributaries on relatively short timescales (10s to 100s of years) during peak flow events. Additional sediment in the tributaries accumulates above sharp bends upstream of their confluence with the Snake River; these materials likely are mobilized only during extreme events on longer timescales (100s to 1000s of years). Steep hillslopes (62% have a slope greater than 40 degrees) have also produced sediment directly to the mainstem over the last 1,000 years, as well as over longer geologic timescales. Rock varnish confirms that episodic small to catastrophic massive landslides continue to occur along both sides of Hells Canyon.

Riverbed material does not appear to produce substantial volumes of sediment and is not extensively reworked during peak events because the channel is largely a stable, armored system. Based on weathering patterns on the armored gravel bars, these features appear to have been armored for a significant period of time (on the order of 100s to 1000s of years), and not as a result of the HCC.

The very large transport capacity of the mainstem appears to have been so effective that the enormous volumes of sediment produced by local sources have not been sufficient to preclude rapid downcutting, as is evidenced by the very existence of the young, narrow, steep canyon. Thus, the mainstem channel has apparently been generally sediment deficient with respect to the local supply of sediment over a long geologic timescale, independent from the HCC.

Although stream flows and sediment dynamics influence geomorphology in Hells Canyon, the Snake River in this reach is controlled, in large part, by the imprint of its complex geologic history. Most hillslope, valley, and channel morphology features appear to be relic features associated with pre-basin regulation hydrologic and geologic events (i.e., prior to the HCC, as well as prior to the numerous other regulation projects in the upper Snake River Basin). Although pre-settlement and pre-regulation conditions are not well known, the weight of evidence strongly suggests that Hells Canyon has been a largely static river system for at least 1,000 years, if not longer.

In this chapter, stream flows reflecting pre- and post-regulation conditions in Hells Canyon are compared. This is followed by an analysis of the morphology in the canyon. Finally, a detailed discussion of sediment dynamics downstream of the HCC is presented. Figure 5.1 provides a close-up base map for the canyon below the HCC, along with local tributary names and mainstem river miles.

5.1. Stream Flows

5.1.1. Outflows to Hells Canyon

This section summarizes the hydraulic discharges from HCD. A complete hydrology and hydraulic modeling report (Parkinson et al. 2001) has been prepared to summarize the existing hydrologic conditions, as well as potential changes to the hydrograph resulting from different HCC operating scenarios.

Immediately downstream of HCD (USGS 13290450), the mean daily flow has been approximately 20,875 cfs since 1965 (Figure 4.2; USGS 2001b). The HCC augments downstream flows by season depending on flood control, activity of resident fish, hydropower, and recreational operational constraints (IPC 2000a). Releases from Brownlee Reservoir, which is the major storage reservoir of the HCC, maintain flows below 9,000 to 13,000 cfs between mid-October and mid-December for spawning. Downstream flows are maintained above spawning flows until smolts emerge in May or June (IPC 2000a). Downstream diurnal stage fluctuations during the winter and summer of more than 2 vertical feet are caused by power peaking operations at the HCC (USACE 1999).

Between HCD (RM 247) and the Imhana River (RM 192), numerous local tributaries contribute to flow in the Hells Canyon reach, particularly during the spring snowmelt runoff period. Although the local tributaries do not have long-term individual gage records, IPC is monitoring discharge in these tributaries. The total combined flow can be estimated based on available USGS gage data. Between the Hells Canyon (20,875 cfs) and Anatone (36,635 cfs), flow increases by 15,760 cfs. If the combined inflows from the Imnaha, Salmon, and Grande Ronde rivers (14,885 cfs) are subtracted from this volume, then the local tributaries contribute a combined mean annual discharge of 875 cfs to the mainstem.

Downstream of the Imnaha River confluence (RM 192), the Imnaha, Salmon, and Grande Ronde rivers contribute a combined 14,885 cfs to the mainstem Snake River (Figure 4.13; USGS 2001b). The Salmon River provides the largest contribution to this input; average annual inflows of 11,270 cfs represent a more than 50% increase in discharge to mainstem flows. This is primarily a result of the large drainage area associated with the Salmon River (36,260 m^2 [14,000 mi^2]). Smaller tributaries also add to steam flows downstream of the major tributaries, but these discharges are not substantial. Downstream of the Grande Ronde the mainstem flows toward Lewiston, Idaho. Based on data collected between 1958 and 2000, the annual mean discharge at Anatone, Washington, is 36,635 cfs (Figure 4.2; USGS 2001).

5.1.2. Comparison of Pre- and Post-HCC Stream Flows

Although the HCC has undoubtedly altered the natural, unregulated hydrograph in Hells Canyon, these changes are not as large as one would expect initially given the magnitude of the flows in the mainstem Snake River and the areal extent of the HCC. As summarized previously, the HCC represents less than 14% of the total storage capacity in the Snake River Basin and is capable of storing approximately 11% of the average annual flow (13 million acre-feet at Weiser). This is markedly different from other large systems, such as the Grand Canyon, where Glen Canyon Dam alone is capable of storing 2.3 years of the Colorado River's annual flow (Collier et al. 1996). The low ratio of reservoir storage capacity to basin runoff indicates that the HCC has relatively little ability to regulate floods because the reservoirs will either be full or fill rapidly. Furthermore, pre- and post-regulation releases are not of sufficient magnitude to mobilize the erosion-resistant geomorphic features in Hells Canyon that exert significant hydraulic control on the river (discussed in Section 5.2.). A specific discussion of the effects of flows on terraces in Hells Canyon is provided in Parkinson et al. (2003).

5.1.2.1. Grams Analysis

As part of the analysis of sandbars in Hells Canyon, Grams (1991) evaluated the pre-and post-hydrographs just below Hells Canyon. He concluded that the pre-HCC 10-year flood was 75,000 cfs and post-HCC flood was about 78,000 cfs. Grams also concluded that the mean annual flood (with a recurrence interval of 2.3 years) is about 50,000 cfs for both time periods (Grams 1991). Grams also concluded that variations on a weekly or daily basis are greater than annual variations because of diurnal and weekly fluctuations in response to electricity demands and additional downstream navigational requirements. Additionally, dimensionless duration curves (normalized for varying mean annual flow) show that although there has been little change in the shape of the duration curves, the HCC has slightly lowered both the peak and low flows (Grams 1991). Specifically, Grams concluded that the HCC has removed the spikes of peak flows above the 30,000 cfs level, which is the maximum power generating capacity of the power plant at HCD (Parkinson et al. 2003).

5.1.2.2. IPC Analysis

These conclusions are consistent with other analyses that have been conducted by IPC on the HCC outflows using the Indicators of Hydrologic Alteration (IHA) software. The IHA software was developed by The Nature Conservancy as a tool for characterizing hydrologic regimes and to analyze changes in those characteristics over time or between data sets (Richter 2001; Richter at el. 1996). Current Operations and Run of River Operations were the two operational scenarios compared using the IHA method. Current Operations are defined as HCC operations under the most recent guidelines and regulations; these data were modeled data produced by CHEOPS. The Run of River Operations simulate pre-dam conditions where inflow to Brownlee Reservoir plus tributary inflow of Wildhorse River and Pine Creek into the HCC equals outflow. Under the Run of River Operations, the reservoirs would be held at full capacity, and all water that entered the system would be passed directly through the HCC. A data set consisting of 72 years (1928-1999) of daily discharge values from HCD outflow was evaluated for each scenario[1].

As shown in Table 5.1, the mean annual flow for the Current Scenario was 19,925 cfs, while the mean flow for the Run of River Scenario was 19,909 cfs. As expected, the presence of the dams does not substantially impact the mean annual flow but shapes it throughout the year. The monthly mean discharges for the entire 72-year period are fairly similar. Only three months (May, September, and November) have large differences. In May, current operations reduce the mean discharge by about 6,400 cfs. As expected in a reservoir used for flood control, the high peaks are reduced. The increase of flow in September under Current Operations is also as expected in a system used for hydropower; low flow periods are increased for power production. The third difference is a reduction in mean discharge in November of about 5,200 cfs. This is an artifact of holding the flows low in the fall for Fall Chinook Salmon spawning so that no spawning occurs in areas that will be dewatered (dried out) in the following spring. This lower flow typically starts in October and extends into December; however, it appears that this operation does not have a substantial impact on the flows in these months except for November.

Minimum flows on a short term basis (i.e. 7-day minimums or less) are increased by current operations, but maximum flows are not substantially affected. Current Operations appear to delay the date of both the high flow in the spring and the low flow in the fall by about 16 days and 47 days, respectively. The rise and fall rates (cfs/day) are both lower in the current operations. This may be attributed to having more control over ramp rates with current operations, which minimize the magnitude of the flow event entering the Snake River below the HCC. The number of flow reversals is fewer under the current operations as well.

Because the values that were input into the model are daily average discharge values, the flow fluctuations within the day are averaged out. The use of daily average discharge limits the analysis of flow fluctuations to daily values. Thus, the diurnal changes that take place during the Current Scenario become absorbed into the daily average discharge value, which was comparable to the daily average discharge value obtained from the Run of River Scenario.

5.2. Morphology

In contrast to the reach upstream of HCD, Hells Canyon is characterized by deep and narrow V-shaped valley canyon entrenched in erosion-resistant basalt and metamorphic bedrock. A

[1] These two data sets were imported into the IHA software, and a trend analysis was performed on each set. Parametric statistics (mean/standard deviation) were used to define the boundaries of the statistical analysis. This analysis produced mean discharges for each month, maximum and minimum daily mean discharge values, pulse counts, as well as the number of flow reversals that occurred. The high and low flow pulse thresholds were defined as the mean plus or minus one standard deviation. The number of reversals refers to the number of times that the hydrograph switched from a rising to a falling condition or vice versa.

geomorphic classification conducted by Simons and Associates (2001) indicated that this reach of the river could be classified as an F1-type stream within the Rosgen (1996) framework:

> *"The F1 stream type is an entrenched, meandering, high width/depth ratio channel that is deeply incised in valleys that are structurally controlled with bedrock...Side slopes of the F1 stream types are often vertical and confine the river laterally for great distances...The dominant channel materials are principally bedrock, with boulders, cobble, and gravel present in fewer quantities. The F1 stream type has a relatively low to moderate sinuosity and low meander width ratios due to the degree of natural entrenchment and lateral containment. The "top of banks" of this stream type cannot be reached by floods that may be developed with the modern-day climate. The F1 stream type typically exhibits low sediment deposition, due to the low sediment supply from the relatively stable bed and banks. These streams are considered very stable stream types due to the resistant nature of their channel materials, and basically have not changed or significantly adjusted in modern times."*

Within this framework, Simons and Associates (2001) concluded that is possible that certain reaches of Hells Canyon could be classified as F2 or F3 (with the higher numbers indicating progressively smaller sediment sizes); however, the dominance of bedrock or large colluvial boulders indicates that the F1 classification is most appropriate for the overall reach of river.

The floodplain is extremely limited in occurrence and extent because of the combined effects of progressive downward scour and gradual geologic uplifting of the mountainous region (see Chapter 1). Therefore, interaction between the river and its bed and banks is largely limited to near-river areas that can be mobilized by the flow, such as bars, islands, terraces, and fans. Although elevated Bonneville Flood terraces are typically located up to tens of meters above the river channel where the river valley broadens or tributaries join the Snake River (Vallier 1998), the Bonneville terraces are functionally disconnected from contemporary Snake River flows because the flood flows calculated by O'Connor (1993) dwarfed those of any contemporary floods.

To provide a complete picture of the valley and channel morphology, this section provides a summary of nearshore characterization study (Section 5.2.1.) and also includes a detailed channel morphology analysis (Section 5.2.2.).

5.2.1. Nearshore Characterization

Detailed analysis of the nearshore geomorphology confirms that the Hells Canyon reach of the Snake River consists of a complex integration of numerous geomorphic features and characteristics (Simons and Associates 2001). A process-based classification approach, such as the one presented by Montgomery and Buffington (1998), was used to address important geomorphic processes that created the various canyon and river features and adapted this to the specific conditions encountered in Hells Canyon. The study focused on a detailed breakdown of *features*, *material types*, and *sources* of material to create a comprehensive description of the Snake River's character as it has developed over time and as it changes along its length. Because virtually all the geomorphic features located above the nearshore environment can be classified as hillslope, the geomorphic classification approach adopted in the study focuses on the nearshore environment. The study is presented in its entirety in Appendix C.

Table 5.2 summarizes the distribution of basic nearshore geomorphic features by 3.2-km (2-mi) river segments. Hillslope or canyon wall is the most dominant feature and is distributed throughout the length of the reach. While there are a few segments with little or no hillslope, hillslope along the 3.2-km (2-mi) segments comprised an average of 54% for both sides of the river, with the maximum hillslope per segment as high as 98 to 100%. On average, 18 to 20% of each 3.2-km (2-mi) segment is a bar, while the maximum amount of bar feature per study segment is 90 to 92%. Bar occurrence generally increases from the upstream end of the reach near HCD (RM 247) to the downstream end of the reach near Asotin, Washington (RM 146). In the reaches where hillslopes become less dominant, bars replace them to a significant degree.

Fans are relatively evenly distributed through the reach, though some 3.2-km (2-mi) segments have no or a relatively sparse amount of fans. Fans average 18% for both sides of the river; the maximum amount of fans per segment is 62 to 96%. The greatest concentration of terraces is located in the upper end of the reach from about RM 239 to RM 219. In the middle to downstream end of the reach, they occur only sporadically. These channel features are discussed in more detail in the following section.

5.2.2. Channel Morphology

Channel morphology and associated physical processes can be examined at a variety of spatial scales ranging from physiographic provinces to channel units (individual pools, bars, riffles, etc.) (Figure 5.2). The Hells Canyon reach of the Snake River is narrowly confined by valley walls and directly coupled to hillslope processes and sediment inputs (Photo 7). Due to the strong valley-wall confinement, the river lacks the floodplain morphology and alluvial character typical of other lowland rivers of comparable gradient and drainage area. Much of the river morphology is forced by large-scale geologic and geomorphic controls that significantly reduce the range of fluvial processes and types of channel adjustment found in other lowland alluvial rivers. Channel processes and morphology at the valley segment and reach scales within Hells Canyon are examined in the following sections.

5.2.2.1. Valley Segment Morphology

At the broadest scale within Hells Canyon the river can be divided into three valley segments (upper, middle, and lower) defined by differences in channel slope (Figure 5.3). The average bed slope of the three valley segments systematically decreases from about 0.002 to 0.0007 in the downstream direction. The break in slope separating the upper and middle valley segments occurs near RM 229 and may be associated with a fault zone in the vicinity of Pine Bar (Figure 1.5, “hooked” fault below Sheep Creek). The division between the middle and lower valley segments occurs near RM 186 and is associated with a fault zone located just downstream of the Salmon River confluence (Figure 1.5). In addition to structural controls, the change in channel gradient between the middle and lower segments may be due in part to contribution of flows from two major tributaries (Salmon and Grande Ronde rivers) and to a lithologic transition from the Seven Devils metamorphics to the relatively less resistant Grande Ronde basalts (Figure 1.5;

Photo 7: Hells Canyon downstream of the confluence with Bernard Creek (RM 235). Note the supply of sediment from the adjacent hillslopes and the structural controls on the channel width.

Chapter 1). Both of these factors may cause greater erosion within the lower valley segment and a reduced channel gradient.

All three valley segments are narrowly confined by valley walls, with little to no floodplain development. Channel confinement is typically defined as the ratio of the bankfull channel width to the valley (floodplain) width (Montgomery and Buffington 1997; 1998). Using this definition, much of the river would have a confinement approaching 1, which is an extreme and rare value for rivers of similar gradient. Lowland rivers of comparable channel slope and drainage area typically occupy broad alluvial valleys that are characterized by extensive floodplains across which the river actively migrates. However, the close proximity of valley walls in Hells Canyon precludes any significant floodplain processes and the development of a typical bankfull morphology (Photo 8).

Channel confinement affects the relationship between discharge, channel hydraulics, and sediment transport. In unconfined rivers, the floodplain provides a relief valve for extreme discharge events. In general, as the river stage rises there is a direct correlation between discharge, shear stress and sediment transport up to the point that the river overtops its banks and spills out across the floodplain. The floodplain relief valve causes the shear stress and sediment transport within the river to level off (or at least increase at a much slower rate) as discharge continues to rise above the bankfull level. In contrast, confined rivers lack this hydraulic relief valve and increased discharges are directly translated into greater shear stresses and greater potential sediment transport.

Photo 8: Hells Canyon between Tin Shed and Big Canyon (RM 213). Note the vertical canyon walls and the strong structural controls on the channel.

Despite the high degree of channel confinement within Hells Canyon, the effects of increasing discharge are modulated by local variations in both the channel width and the slope of the valley walls. This variability is illustrated by comparing the flow width of the 1.5-year discharge (comparable to the recurrence interval of a bankfull flow [Leopold et al. 1964]) to that of the 100-year flow ($W_{1.5}/W_{100}$), which provides a hydrologic definition of confinement (Figure 5.4). Values of $W_{1.5}/W_{100}$ range from less than 0.5 to almost 1, illustrating the local variation in channel width and/or valley-wall slope (Figure 5.4). Nevertheless, the average channel confinement is quite high and shows only a slight decrease in the downstream direction.

In this analysis, flow widths were determined from the 1-D hydrodynamic model MIKE 11 (DHI 2000) using topographic cross sections spaced at intervals of about 0.2 km (0.12 mi). Channel and canyon topography were determined from a combination of SONAR and LIDAR surveys (Butler 2001). Discharges used in the model are presented in Table 5.3 and were derived from a Log-Pearson Type III analysis of annual peak flows (USGS 1982; Thomas et al. 1998) adjusted for regional skew values (Kjelstrom and Moffatt 1981). Inputs from the major tributaries (Imnaha, Salmon, and Grande Ronde rivers) were included in the analysis, while discharge contributions from the smaller tributaries along Hells Canyon were considered negligible.

5.2.2.2. Reach-scale Morphology

5.2.2.2.1. Sinuosity and Confinement

The three valley segments can be further divided into channel reaches based on qualitative differences in sinuosity and type of confinement (Figure 5.5; Table 5.4). Sinuosity varies from low (nearly straight sections, e.g., Reaches 1-3) to moderate (short-wavelength bends in Reach 4) or long wavelength bends (collectively observed in Reaches 5-9). Although all of the reaches are confined, the type of confinement varies from bedrock walls and talus slopes to mixtures of bedrock and alluvial terraces (Bonneville Flood deposits, landslide backwater deposits, relict alluvial fans, or relict bars).

5.2.2.2.2. Reach Type

At the reach scale, the river exhibits a regular pool-riffle morphology between HCD (RM 247) and Asotin, Washington (RM 146). Figure 5.6 shows the typical repeating sequence of pools and riffles observed along the thalweg profile. The frequency (or spacing) of pools varies from reach to reach but typically has a value of 4 to 9 channel widths, with a median value of about 6 (Figure 5.7). This pool spacing is within the range typically reported for pool-riffle channels. Average pool spacing in self-formed pool-riffle channels (those not forced by external controls) is typically 5 to 7 channel widths (Leopold et al. 1964; Keller and Melhorn 1978; Thompson 2001), but can be as low as 0.1 channel widths in obstruction-forced pool-riffle channels (Montgomery et al. 1995; Buffington et al. in press a).

5.2.2.2.3. Channel Units

Reach morphology was further classified by mapping channel units (individual pools, bars, riffles, etc.) (e.g., Bisson et al. 1982; Woodsmith and Buffington 1996) from 1:8400 aerial photographs (USACE 1996; IPC 1997) and from topographic maps drawn with 2-meter (6.6 ft) contour intervals. As discussed above, channel topography was determined from SONAR and LIDAR surveys (Butler 2001). The channel-unit maps are shown in Appendix D (Panels 1-12), with descriptions of each channel-unit type provided in Table 5.5.

Channel units were distinguished based on morphologic criteria and, for the most part, are not stage-dependent. For example, some channel-unit classifications distinguish different types of shallows (runs, riffles, rapids, cascades) based on flow characteristics (tranquil, rough, tumbling, etc.), which can make the classification stage-dependent. Flow over coarse bed material may appear rough at low discharges, but tranquil at higher flow as the stage increases and moves away from coarse bed material. In this analysis, rapids were the only channel unit identified from flow characteristics.

Specifically, morphologic and hydraulic were the two types of rapids that were identified. Morphologic rapids are defined as having relatively steep channel slopes, boulder bed material, and macroscale turbulence (white water). The boulders are locally supplied by rock falls, landslides, or tributary debris flows. In contrast, hydraulic rapids are due to local flow convergence that produces standing waves and turbulent, tumbling flow, but may or may not contain boulder-sized bed material. Hydraulic rapids most commonly occurred due to topographic steering and shoaling over debris-flow fans as the flow accelerates into an associated pool scoured along the fan margin.

Geomorphic features near the channel also were identified and mapped where they provided insight regarding channel morphology and processes. These features included debris fans, talus slopes, landslides, and alluvial terraces (Bonneville Flood deposits, landslide backwater deposits, relict alluvial fans, or relict bars) (Appendix D, Panels 1-12). In particular, debris-flow fans (predominantly emanating from high-gradient tributaries) are ubiquitous in the Hells Canyon reach and play a dominant role in channel morphology and fluvial processes. Over 420 debris fans enter the channel within the Hells Canyon study area (roughly 3 fans per km), with 67% of these influencing channel morphology.

As described in recent studies of the Grand Canyon (Schmidt and Rubin 1995; Andrews et al. 1999; Webb et al. 1999), debris fans create characteristic channel units and associated process domains (Figure 5.8). As a debris fan grows and builds out into the channel it creates a flow obstruction that constricts the flow and forces the flow to converge opposite the fan. The constriction increases flow velocity and shear stress, causing pool scour opposite the fan. Downstream of the fan a large eddy is created in the lee of the fan, generating a backwater (depositional) environment in which eddy bars (sand beaches) may be formed. Bed material scoured from the pool may be entrained by secondary currents and deposited along the channel margins or in the eddy zone downstream of the fan. Alternatively, bed material scoured from the pool may be deposited in a downstream bar where the flow diverges as it exits the pool, as typically occurs in an alternate bar (pool-riffle) morphology (e.g., Montgomery and Buffington 1997). Shoaling of the flow over the fan and convergence into the pool also commonly creates hydraulic rapids.

Alternatively, morphologic rapids may occur where episodic debris flows and tributary floods deposit boulders in the mainstem channel. The channel gradient of the mainstem river is typically insufficient to mobilize these boulders, so that over time extensive boulder rapids can develop adjacent to debris fans. Examples of these morphologic features within Hells Canyon can be seen in the channel-unit maps (Appendix D, Panels 1-12).

Debris fans are formed by several geomorphic processes in the study area: 1) tributary floods and debris flows in channeled valleys, 2) debris flows and snow avalanches on hillslopes and unchanneled valleys, 3) landslides, and 4) gravity-driven transport of talus (rock fragments). Tributary fans are the most dynamic because of the relatively high frequency of fluvial sediment transport events (on the order of 0.1-10 years) due to relatively large contributing areas that support frequent hydraulic discharge. Landslides, snow avalanches, and debris flows occur less frequently (on the order of 10-10,000 years), but may carry large amounts of sediment when they do occur. For example, the Rush Creek landslide was a rare but catastrophic event that deposited roughly 120 meters (400 ft) of sediment in the mainstem Snake River and temporarily dammed the river (Vallier 1998). The tributary fans represent both frequent fluvial events and more episodic debris flows, with the debris-flow deposits reworked by fluvial processes both within the tributaries and in the mainstem river (e.g., Webb et al. 1999). Talus fans are the least dynamic and slowest growing features and are dependent on processes of creep, rock avalanche, frost heave, and shaking from earthquakes.

5.2.2.2.4. Pools

Approximately 91% of the 175 pools inventoried in the channel-unit mapping are forced either by debris fans or bedrock (Figure 5.9). Like debris fans, bedrock projections locally create flow obstructions that cause flow convergence and turbulent pool scour. Bedrock constrictions caused by narrowing of the channel walls also force flow convergence that may scour a pool. Less than 8% of the pools inventoried were self-formed (not forced by external controls).

Distributions of residual pool depth (which is the difference between pool bottom and downstream riffle crest [Bathurst 1981; Lisle 1987]) do not show any significant differences across the different pool types (Figure 5.10), nor are there any significant trends in residual depth along the length of the study area (Figure 5.11). However, the range of residual depths for pools formed by tributary fans does appear to increase in the downstream direction. Figure 5.11 also shows that pools formed by talus fans are restricted to the uppermost channel reaches (just below HCD), while bedrock-forced pools are more common in the lower reaches. Pools forced by tributary fans occur throughout the study area. Figure 5.11 also suggests that there is a lower limit for residual pool depth of about 4 meters (13 ft).

5.2.2.3. Hydraulic Geometry

Results from the 1-D MIKE 11 hydrodynamic model (DHI 2000) were used to predict downstream hydraulic geometry relationships (Leopold et al. 1964) between HCD (RM 247) and Asotin, Washington (RM 146). Values of average velocity (u), cross-sectional area (A), top width (W), and depth (D) were predicted for the $Q_{1.5}$ flow and plotted versus discharge. As discussed earlier, the hydrodynamic model is driven by four discharge inputs (Table 5.3). Consequently, the hydraulic geometry relationships were developed for four downstream discharges (Snake River at HCD, plus the Imnaha, plus the Salmon, plus the Grande Ronde), with the above channel characteristics averaged over the reaches between input nodes.

Figure 5.12 shows the predicted downstream changes in channel characteristics for the $Q_{1.5}$ flow. Predicted values of width and depth agree with the range of values reported for other natural channels with bankfull discharges comparable to the $Q_{1.5}$ values used here (Figure 8 in Ferguson [1986]). However, the predicted exponents for the width, depth, and velocity functions differ from other values reported for alluvial rivers. In particular, the predicted exponent for the width relationship is higher than typically reported (0.80 versus 0.40-0.50), while the exponent for the depth function is considerably lower than typically reported (0.05 versus 0.30-0.45) (Leopold et al. 1964). These differences may be driven by the high degree of confinement and non-alluvial nature of the Hells Canyon study area. It is interesting to note that recent investigations by Montgomery and Gran (2001) indicate that typical hydraulic geometry relationships reported for alluvial channels should also apply to bedrock reaches (which are also non-alluvial channels with high confinement values).

The individual predicted values of downstream channel characteristics (rather than the reach-average values used in Figure 5.12) reveal interesting downstream spatial controls on channel form (Figure 5.13). In particular, downstream values of channel width and flow area exhibit subtle wave-like oscillations about constant average values over the first 88 km (55 mi) downstream of HCD (RM 248-RM 192). The lack of a systematic increase in channel width in the first 88 km (55 mi) downstream of HCD suggests that tributary discharge inputs in this section of the river have a negligible effect on channel dimensions. The cause for the wave-like oscillations in channel width in the first 88 km (55 mi) is uncertain, but may be a legacy of paleoflood hydraulics (e.g., the Bonneville Flood). Undulating channel width is common in canyon rivers and is believed to provide hydraulic roughness that minimizes downstream energy expenditure in a fashion analogous to bedform topography (Wohl et al. 1999).

Below where the Imnaha and Salmon rivers discharge into the mainstem (at about 88 km (55 mi) downstream of HCD), width and area rapidly increase and show a direct relationship with increasing downstream distance (and thus increasing discharge). The increase in channel width and flow area corresponds with the beginning of the tributary inputs at the Imnaha and Salmon rivers and also corresponds with a lithologic change from the Seven Devils metamorphics to the relatively less resistant Grande Ronde basalts (Chapter 1). Consequently, the observed increases in channel width and flow area likely result from the combined effects of tributary discharge inputs and the transition to a relatively weaker lithology that may allow more rapid lateral erosion.

5.2.2.4. Regime Diagrams

Regime diagrams describe the physical conditions and specific combination of flow and sediment transport characteristics that give rise to a given morphologic state. Parker (1990) developed a particularly well-formulated regime diagram that links equations for hydraulic discharge, bedload transport, and channel characteristics (grain size, flow depth, and slope). The Parker framework relates dimensionless bedload transport rate (q_b*) to dimensionless hydraulic discharge per unit width (q*). The dimensionless bedload transport rate is defined here from the Meyer-Peter and Müller (1948) equation as

$$q_b^* = 8\,(\tau^*_{50s} - \tau^*_{c50s})^{1.5} \qquad (5\text{-}1)$$

where τ^*_{50s} is the dimensionless shear stress of the median surface grain size (d_{50s}) determined from the Shields (1936) equation

$$\tau^*_{50s} = \frac{\tau_0}{(\rho_s - \rho)\, g\, d_{50s}} \qquad (5\text{-}2)$$

In Equation 5-2, ρ_s and ρ are the sediment and fluid densities (set equal to 2800 and 1000 kg/m^3, respectively), g is the gravitational constant, and τ_0 is the total boundary shear stress determined as a depth-slope product ($\tau_0=\rho gRS$, where R is the hydraulic radius and S is the energy slope). τ^*_{c50s} is the critical dimensionless shear stress for incipient motion of d_{50s} and is set equal to 0.03. Reported values of τ^*_{c50s} typically range from 0.03-0.09 for rough turbulent flow (Buffington and Montgomery 1997). Consequently, setting τ^*_{c50s} equal to 0.03 provides a conservative estimate of particle mobility.

The dimensionless specific hydraulic discharge is defined as

$$q^* = \frac{<u>D}{\sqrt{\boldsymbol{R} g d_{50_s}}\, d_{50_s}} \qquad (5\text{-}3)$$

where D is flow depth, $\boldsymbol{R}$ is the submerged specific gravity of sediment [$(\rho_s\text{-}\rho)/\rho$, set equal to 1.80], and $<u>$ is the vertically-averaged velocity determined from the law of the wall (Keulegan 1938):

$$<u> = \frac{u^*}{\kappa} \ln\left(\frac{0.368\, D}{z_0}\right) \qquad (5\text{-}4)$$

In Equation 5-4, u^* is the shear velocity ($\sqrt{\tau_0/\rho}$), κ is von Karman's constant [0.408, Long et al. (1993)] and z_0 is the height above the bed where the velocity profile goes to zero. The Whiting and Dietrich (1990) approximation is used to define z_0 as $0.1d_{84}$, where d_{84} is the surface grain size for which 84% of the sizes are smaller. Combining Equations 5-3 and 5-4 and defining $D^* = D/d_{50s}$ yields

$$q^* = \frac{\sqrt{\tau^*_{50s}}}{\kappa} D^* \ln\left(\frac{3.68\, D}{d_{84}}\right) \qquad (5\text{-}5)$$

Measured grain sizes and hydraulic variables predicted from MIKE 11 for the $Q_{1.5}$ flow were used to calculate values of q^* and q_b^* and were compared to data from other channel types (Figure 5.14). Despite the non-alluvial character and strong external controls on channel morphology, the pool-riffle data from Hells Canyon plot alongside data from pool-riffle channels in floodplain rivers of North America and Britain. This result supports arguments made by Buffington et al. (in press a, b) that different channel morphologies arise from mutual adjustment of channel characteristics (width, grain size, bed slope, etc.) to imposed watershed conditions (hydraulic discharge, sediment supply, valley slope, etc.). Consequently, one would expect that data from a given channel type (e.g., pool-riffle channels) should plot near one another in a regime diagram that relates imposed watershed conditions to channel characteristics.

Nevertheless, the non-alluvial character and external controls imposed on the Hells Canyon reaches limit the range of potential channel adjustments to a given perturbation compared to similar alluvial channel types, as will be discussed further in Chapter 6.

5.3. Sediment Dynamics

IPC has completed extensive study of the sediments in Hells Canyon reach downstream of the HCC. Based on the results of these analyses and the general character of the canyon, this section is divided into two major components: 1) potential local sediment sources to the mainstem and 2) mainstem sediment transport dynamics.

5.3.1. Potential Sediment Sources

The upstream sources of sediment (i.e., sediment sources that would have come from the upper reaches of the Snake River but have been cut off from downstream transport by the numerous storage facilities, including the HCC) were previously addressed in Chapter 4. Sources of local sediment supply to Hells Canyon can be divided into four categories:

1. Local tributaries
2. Adjacent hillslopes
3. Riverbed material
4. Riverbank material

Each of these areas is discussed in more detail below (see Figure 5.1 for specific tributary locations).

5.3.1.1. Local Tributaries

Within Hells Canyon, local tributaries are considered sediment production zones in Schumm's general classification scheme (Figure 4.3). Upstream of the Imnaha and Salmon rivers (RM 192 and RM 188, respectively) more than 30 local tributaries periodically erode and produce sediment (Photo 9). While the small tributaries within Hells Canyon are small in comparison to the larger upstream tributaries, their influence on local processes of sediment production, transport, and deposition is significant because they are linked directly with the mainstem Snake River in this reach. These sources are also significant relative to the reduced supply from the larger upstream tributaries that have been cut off by other regulation projects. Supplies of sediment from the local tributaries in this reach have been evaluated over different timescales.

5.3.1.1.1. Short-term Quantitative Sediment Yield Estimates

The potential sediment yield from local tributaries between the HCD (RM 247) and the Salmon River (RM 188) was estimated by developing a transport rating curve (flow versus sediment transport curve) and calculating sediment transport at several different flows (Parkinson et al. 2003). This methodology is based on transport capacity and provides an indication of potential transport under current flows. The estimates are considered conservative

Photo 9: Hells Canyon at Deep Creek (RM 199). Note the fresh 1997 tributary debris deposits at the mouth of the creek on the left side of the channel.

because they are based on discharges up to a 1% recurrence interval. Thus, these estimates do not incorporate extreme events that have less than a 1% chance of occurring on timescales of 100s to 1000s of years. Additional transport capacity is undoubtedly available for these larger events (i.e., 500- or 1000-year return events) but it cannot be estimated well using current flow records with relatively short periods of record (50 to 100 years).

The tributaries with calculated sediment supply located between the HCC and the Salmon River (not including the Imnaha River drainage) account for approximately 901 km^2 (348 mi^2) of a total watershed area of 1,398 km^2 (540 mi^2). The average sediment yield from these calculated tributaries was applied to the remaining area (497 km^2 [192 mi^2], or 36%), which produces a total sediment supply of 8.60 million tons per year (Parkinson et al. 2003). On a strictly volumetric basis, this estimate is about 3 times higher than the annual supply of sediment that has been retained by Brownlee Reservoir (2.78 million tons) since 1958 (Chapter 4). In addition, these local sediments include coarser materials >0.062 mm (>0.002 inch) that contribute more significantly to important channel features such as sandbars and spawning sites.

Photo 10: Hells Canyon upstream of Temperance Creek Fan (RM 224). Note how the tributary fan on the left side of the channel realigned the river course and caused the river to cut into the Bonneville Flood terrace on the right side of the channel.

The average sediment yield of 15,900 tons/mi^2 per year is in the upper range of values found in literature (Parkinson et al. 2003). However, given the characteristics of the tributaries in Hells Canyon including steep slopes, relatively small drainage areas, and limited ground cover resulting from arid conditions, high sediment yields are expected. Furthermore, it is likely that sediment yields prior to the last 1,000 years were even larger than current estimates, given the significantly larger discharges associated with these conditions (Chapter 1 and Section 3.2.). Photo 10 demonstrates that the volume of sediments shed from these tributaries is sufficient in some locations to alter the course of the mainstem river.

Of the 8.60 million tons, sand and spawning-size gravels, respectively, are 1.44 and 4.14 million tons per year (Parkinson et al. 2003). As discussed in Section 5.3.2., these sediment sizes are the most useful in maintaining channel features such as spawning sites and sandbars. Also, on a spatial scale, the two largest tributary suppliers of sediment (Granite Creek and Sheep Creek) are located in the upper part of Hells Canyon above most of the sand beaches and spawning areas identified as areas of particular concern (Parkinson et al. 2003). The elevated sediment yields in Granite Creek are likely due to the glaciated headwater area in this tributary. This means that high-producing tributaries are located upstream of areas that benefit most from sediment supply (e.g., sandbars and spawning sites).

5.3.1.1.2. Long-term Sediment Yield Considerations

In general, short-term sediment yield estimates based on conventional sediment yield measurements tend to greatly underestimate the long-term supply of sediment. For example, Kirchner et al. (2001) recently conducted a sediment yield study in mountainous central Idaho that suggests that conventional sediment yield measurements made over decades greatly underestimate (by an average of 17 times too low) the long-term (over millions of years) average rates of sediment delivery.

In Hells Canyon, long-term (100s to 1000s of years) sediment yields in the tributaries are also likely larger than the short-term yield estimates. However, several factors appear to somewhat limit the volume of sediments that are mobilized from the tributaries to the mainstem channel. Annual peak flows mobilize sediments in the upper portions of the tributaries, but most of this sediment is prevented from reaching the mainstem until more extreme flows are episodically available. Extraordinary streamflow events such as rain-on-snow conditions appear to create sufficient stream flow to mobilize a considerable amount of accumulated sediments from storage into and through the narrow creek channels leading to the Snake River. An example of this occurred in January 1997, when debris flows mobilized substantial volumes of tributary sediment throughout the Hells Canyon reach. Remnants of one of these debris flows were observed during field visits to Wolf Creek in 2001. In the four years since the 1997 debris flow, the creek has cut vertically down approximately 3 m (10 ft) through debris flow sediments, leaving walls of debris on both sides of the creek.

In addition to limited stream flows, the geometry of the tributaries is believed to cause a substantial volume of sediments to be stored. Most of the major tributaries have a very unusual 45- to 90-degree bend within 0.8 km (0.5 mi) of their confluence with the Snake River (Photo 11). The causes of these bends are unknown, although the tributaries may be controlled by a fracture, joint, or other linear structural break in the bedrock in response to rapid downcutting by the Snake River. These tributary bends may also be a result of the reverse in stream flows caused during the capture of the Snake River, as discussed in Chapter 1.

Typically, above this bend the creek floodplain stores a tremendous amount of sediment (Photo 12). However, below the bend the creek and floodplain become increasingly restricted as the creek approaches the confluence with the Snake River. Essentially, below the bend the creek typically forms a chute either between bedrock walls or between a very narrow and vegetated stable floodplain. The sediment accumulation above the bend indicates that transport capacity of the creek is severely impacted by the bend and narrowing of the creek below the bend. The decrease in stream power created by the nearly right angle bend, coupled with the restricted creek channel width below the bend, probably causes the significant sediment accumulation above the bend and lack of transport capacity to move these sediments to the Snake River except under significant episodic flow events.

Photo 11: Temperance Creek (RM 223). Note the old landslide surface on the south (bottom) side of the creek indicating that in the geologic past Temperance Creek did supply sediments to the Snake River. Relatively well-developed vegetation along the entire channel suggest that there has not been a recent episodic movement of sediment into the Snake River from Temperance Creek.

Photo 12. Example of sediment stored in an alluvial debris fan within the Battle Creek drainage near its confluence with the Snake River (RM 240). The adjacent small tributary to Battle Creek also has formed a large alluvial debris fan of actively moving gray sediments into the creek and river.

The presence of debris flow fans in the mainstem channel indicates that episodic, very large rain-on-snow events can produce sufficient surface water flow to move some accumulated sediment around the bend and down the chute below the bend. Alluvial/debris fans formed at the confluence of several tributaries with the Snake River are typically highly variable in size, and their size does not appear to be proportional to their drainage area. For example, the Imnaha River does not have a major alluvial/debris fan compared with other smaller tributaries. This variability in size could be due to a number of factors, including either the tributary bends or different uplift rates throughout different segments of the canyon (discussed in more detail in Section 5.3.2.). The variability in debris fan size also may be because the larger drainage basins probably have sufficient stream power to move the debris out into the Snake River. In contrast, the smaller drainages probably do not have sufficient stream power to move the sediment much beyond the mouth of the tributary. Once the debris fans are deposited, smaller particles are transported downriver while large particles (cobbles and boulders) contribute to, and form, rapids below the tributary confluence (as discussed previously).

Most of the alluvial/debris fans have a varnish coating (suggesting they are at least 1,000 years old, as discussed in Section 5.3.1.2.) and many of them are vegetated. Thus, these debris flows appear to be composed of episodically-mobilized sediments that have been accumulating for a long period of time within the small drainages. An estimate of the volume of sediment that has been mobilized under these conditions cannot be made, but the substantial influence of debris fans to the channel morphology suggests that this volume has historically been considerable over timescales of 100s to 1000s of years, or longer.

5.3.1.2. Hillslopes

Adjacent hillslopes were also evaluated as a second potential source of sediments to the mainstem. The depth and steepness of the current canyon form suggests that hillslope material has been eroding steadily over the last 2 to 6 million years at a rapid rate.

5.3.1.2.1. Short-term Quantitative Sediment Yield Estimates

Quantitative estimates of sediment yield from the hillslopes are difficult to calculate and no literature values for comparable systems are available. If one assumed that the hillslope processes that produce sediment in the local tributaries are similar to the hillslope processes that produce sediment directly to the mainstem, the sediment yield from the local tributaries could be applied to the amount of drainage area that is directly adjacent to the mainstem. A yield of 15,900 tons/mi^2 per year was applied to the 53 mi^2 of hillslopes that drain directly into the mainstem Snake River. This results in an estimate of 0.84 million tons per year (387 acre-feet per year) of hillslope sediments larger than 0.063 mm (0.002 in) that are shed directly into the mainstem channel.

5.3.1.2.2. Long-term Sediment Yield Considerations

The short-term estimate of hillslope sediment yields is a conservative estimate that does not account for longer term, episodic and catastrophic events such as the Rush Creek landslide. Near RM 231, this catastrophic landslide cascaded into the mainstem and temporarily damned the Snake River to a depth of nearly 120 m (400 ft) (Vallier 1998). Thus, on longer time scales (100s to 1000s of years), a considerably larger amount of sediments appear to have been shed. On a qualitative level, slope classifications and varnish interpretation were two lines of evidence used to evaluate long-term hillslope yields.

Slope Classifications

Figure 5.15 shows the five slope classifications for hillside slopes in the Hells Canyon drainage basin (excluding the Imnaha River and Salmon River tributary drainage basins). Table 5.6 summarizes the distribution of hillside surface area within each of the five slope classes. Although a more detailed hillslope analysis by individual slope class is provided in Appendix E, the following discussion provides an overview of how different slope classes may affect sediment delivery to the local tributaries and to the Snake River mainstem channel.

Approximately 20% of the area in the canyon has slopes that are below 30 degrees. These slopes are typically below the angle of repose, which is “maximum angle of slope at which loose, cohesionless material” of similar composition will come to rest (Gray et al. 1974). Slopes flatter than the angle of repose depend primarily on surface runoff to transport sediments to streams. In this case, this slope class is closely tied to the large expanse of relatively flat-lying CRBs in this drainage basin (Table 5.7).

The angle of repose where sediment particles tend to easily move downslope commonly ranges between 33 and 37 degrees. Approximately 18% of the slopes in the canyon fall into the slope classification between 30 and 40 degrees, or near the angle of repose. Most of the surface with this slope category is likely to be loose material of variable particle sizes forming scree slopes and talus piles. Upstream of RM 217, most of these slopes are located in the headwater drainage areas, which supports the idea a substantial portion of sediment is produced and stored in the individual tributaries. Downstream of RM 217, surface areas with this slopes between 30 and 40 degrees cluster somewhat near the river, which suggests that most of these areas are associated with scree slopes and talus piles directly adjacent to the Snake River (Photo 13).

Most of the surface area (62%) in the drainage basin has a hillside slope that is larger 40 degrees and an additional 10% of the surface area has a slope greater than 60 degrees (Table 5.6). Given the nature of the canyon and bedrock walls that line the canyon valley, the dominance of steeper slopes is not too surprising. Slopes greater than the angle of repose are likely to be made up of material that is somewhat durable or cohesive and is transported to streams via rock falls, debris avalanches, and/or landslides. In bedrock-dominated environments such as Hells Canyon, surfaces greater than 40 degrees will tend toward catastrophic failure events.

Frequently, the creeks and streams are adjacent to, but do not flow through, major landslides. This indicates that the creeks precede the landslide and are probably not a significant part of the landslide-forming process. An exception to this may be where there are landslides on the sides of tributary creeks above their confluence with the Snake River (for example, Temperance Creek at RM 223). In much the same way that landslides occur along the sides of the Snake River, the creeks undermine steeply dipping bedrock units and transport the sediments discharged by the landslides.

Photo 13. Lower Dry Gulch (RM 237). Note scree slopes forming the chutes of sediments moving toward the river and stable vegetation along the river edge.

Most of areas with slopes greater than 60 degrees are located adjacent to the Snake River or in the lower parts of individual drainage basins above about RM 217. Indeed, this slope class is associated with the scenic high canyon walls between RM 247 and RM 217. The dominantly steep slopes in Hells Canyon indicate a very young surface that has been exposed to earth surface weathering processes and rapid downcutting by the Snake River for only a very short geologic time.

Rock Varnish

Color aerial photographs taken in August 1997 were used to interpret locations in the Snake River Canyon where significant volumes of sediments are mobilized into the river. Again, although a more detailed varnish analysis is provided in Appendix E, the following discussion provides an overview of how varnish may reflect sediment delivery to the Snake River mainstem channel.

Rock varnish is a orange, yellow-brown, brown, and black coating that ubiquitously occurs on rocks, sediments and soils exposed to the dust, aerosols and other airborne material in the atmosphere (Kinsley 1998). Although rock varnish is found in all climatic conditions and all lithologies, it is most evident in arid to semi-arid environments. Under moist conditions (for example, the black rock varnish on rocks adjacent to the Snake River), it can form in a few decades (Liu and Broecker 2000). In the semi-arid to arid conditions in Hells Canyon, yellow-brown to orange rock varnish probably requires about "3,000 to 5,000 years to form a visually discernible patchy varnish and 10,000 years for a heavily coated varnish." An example of rock varnish in the canyon is shown in Photos 11, 12, and 13.

Rock varnish tonal colors indicate a relative age of sediment disturbance or movement. To be conservative, a surface without varnish tonal color was considered to have been disturbed by either natural or anthropogenic processes in the last 1,000 years. In reality, some of the disturbances could have occurred from slides and slumping sediment movement just before the aerial photograph was taken. Most of the anthropogenic disturbances in Hells Canyon have probably occurred in the last 100 to 200 years based on the literature estimates of the minimum rate of varnish formation (Fleisher et al. 1999, and Liu and Broecker 2000).

Photos 7, 11, 12, and 13 show that the hillslopes in Hells Canyon actively shed sediments over short-term (within the last 1,000 years) and long-term (longer than 1,000 years) timescales. Episodic small to massive landslides have occurred along both sides of Hells Canyon and have obviously produced impressive sediment volumes into the Snake River over recent geologic history. In addition, some landslide surfaces continue to shed sediments into the river. Based on the varnish analysis, sediments have been mobilized at 43 locations within the last 1,000 years, with approximately half of these locations on either side of the canyon.

In the upper reaches of the canyon (RM 247 to about RM 235), mobile sites are about twice as common on the Idaho (eastern) side of the canyon, which is probably related to the metamorphic bedrock along this side of the canyon with slopes exceeding 60 degrees. The largest landslide in this portion of the canyon is the Waterspout landslide at RM 234, which is a major current and historical source of sediment to the river (Photo 14).

Between RM 235 (which corresponds to a major change in lithologies) and RM 217, both sides contribute sediments about evenly. Major features in this reach include Rush Creek landslide (RM 231) and the High Bar and Alum landslides (RM 227), the Temperance Creek debris fan/landslide complex (RM 223; Photos 10 and 11), and an unnamed Pittsburg Landing landslide (RM 215). In addition, Bonneville Flood terrace gravels are being actively eroded into the Snake River. Most of the remaining sediment transport to the Snake River is from numerous scree slopes from both sides of the canyon.

The Snake River makes a significant change from northeasterly flow to northwesterly flow at about RM 217. Between this point and RM 179 the river has numerous actively shedding sediment features, the slight majority of which originate on the Oregon (western) side of the channel. The color tone of the varnish and level of erosion on the landslide surfaces along this segment of the river and a significant amount of erosion (e.g., strongly incised drain lines cutting through the surface) both suggest that these landslides are generally older than other upstream landslides. Structural differences are probably controlling the landslide processes in the area. For example, significant landslides between RM 207 and RM 208 within meta-sedimentary units occur on the Oregon (western) side of the river. The slides form triangular-faceted surfaces on several slopes, which suggests that dip-slope failure caused the landslides.

5.3.1.3. Riverbed Materials

The potential sediment yield from the riverbed material in the mainstem Snake River was also evaluated. For the purposes of this section, bed stability was determined using calculations of incipient motion and cross-section measurements at locations within Hells Canyon (Parkinson

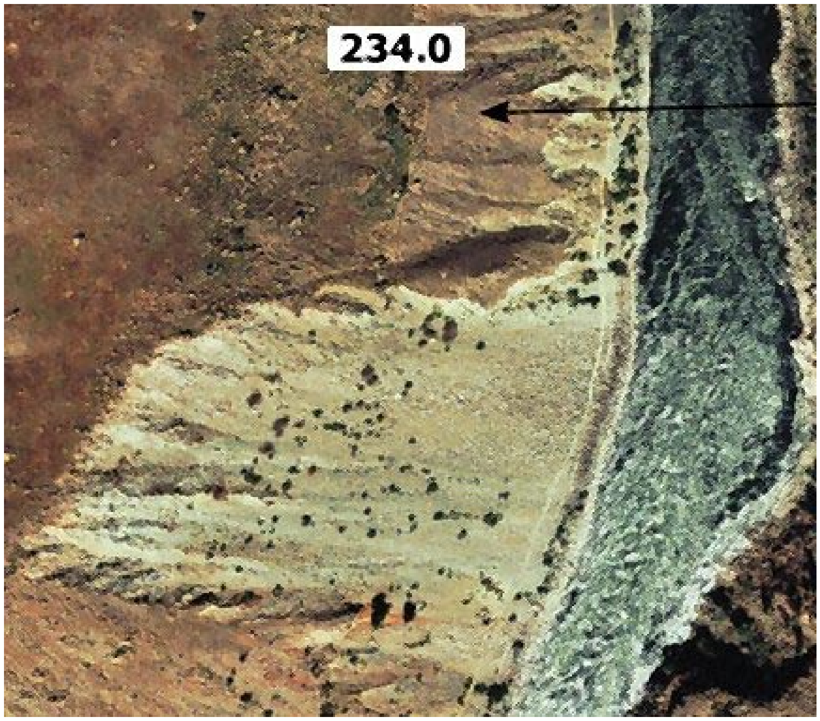

Photo 14: Waterspout Rapids (RM 234). In this case, sediments are actively moving into the river from a slide surface on the left side of the river.

et al. 2003). Additional information on the transport of bed materials that are mobile within the mainstem is provided in Section 5.3.2.

5.3.1.3.1. Bed Mobility

Incipient motion results for the mainstem indicate that isolated pockets of bed material move under the flows currently experienced in this reach, but the majority of the bed appears to be stable (Parkinson et al. 2003). The areas associated with mobile bed material are typically located downstream of a tributary. This suggests that historical large flow events mobilized and deposited sediments from local tributaries in isolated pockets downstream of the parent tributary.

Available grain-size data also can be used to assess the general mobility of the bed and bar sediments[2]. Figure 5.16 compares the Shields (1936) stress for the $Q_{1.5}$ flow to typically reported Shields values for incipient motion (τ^*_{c50s} = 0.03-0.09) (Buffington and Montgomery 1997). The Shields stress for the median surface grain size (τ^*_{50s}) is defined here as

$$\tau^*_{50s} = \frac{\tau_{sf}}{(\rho_s - \rho)\, g\, d_{50s}} \qquad (5\text{-}6)$$

[2] The surface grain size dataset is currently being verified. Minor changes to this dataset may be warranted but are expected to only slightly alter the figures and percentages cited in this report. Conclusions contained herein will not be substantially altered.

where τ_{sf} is the skin friction stress (that portion of the total boundary shear stress acting on the bed and responsible for sediment transport). The skin friction stress is determined iteratively from the Einstein and Barbarossa (1952) method using a velocity profile defined from the law of the wall (Keulegan 1938)

$$\tau_{sf} = \rho \, g \, R_{sf} \, S = \rho \left[\overline{u} \, \kappa \left(ln \frac{0.368 \, R_{sf}}{z_0} \right)^{-1} \right]^2 \qquad (5\text{-}7)$$

where R_{sf} is the hydraulic radius pertaining to the skin friction of the bed material, S is the energy slope, $\overline{u}$ is the local average velocity, κ is von Karman's constant [0.408, Long et al. (1993)] and z_0 is the height above the bed where the velocity profile goes to zero. The Whiting and Dietrich (1990) approximation is used to define z_0 as $0.1d_{84}$, where d_{84} is the surface grain size for which 84% of the sizes are smaller. Values of S and $\overline{u}$ were determined from MIKE 11 simulations of the $Q_{1.5}$ flow (see Section 5.2.2.). The value of τ^*_{c50s} was conservatively set to 0.03, which at the lower end of the reported range of 0.03–0.09 for rough, turbulent flow (Buffington and Montgomery 1997). Figure 5.16 shows that 75% (80 of 107 surface samples) sampled bar surfaces are predicted to be immobile at $Q_{1.5}$ (i.e., $\tau^*_{50s} < \tau^*_{c50s}$=0.03).

In addition to the quantitative estimates of bed material mobility, cross-sectional data from the USGS gage below HCD (13290450) between 1968 and 2001 indicate that this section of river has been very stable with no apparent trends in either the vertical or horizontal direction since 1968 (Parkinson et al. 2003). Similarly, more limited data from the USGS gage near the confluence with the Imnaha River at Joseph, Oregon (13209500) suggest that the river has been stable since at least when the HCC was completed (Parkinson et al. 2003). These data indicate that the channel bed appears to contribute only limited volumes of sediment supplies to the system.

It is not surprising that the incipient motion calculations confirm that the channel bed in Hells Canyon is largely stable. Inspection of the gravel bars that are exposed during typical flows indicates that these materials are strongly imbricated and well armored, as discussed in more detail in the following section.

5.3.1.3.2. Grain Sizes and Armoring

Surface grain sizes sampled on bars in the upper and middle valley segments (Parkinson et al. 2003) demonstrate that the bars are predominately composed of gravel- and cobble-sized material (Photos 15 and 16; Figure 5.17). Nine of the 107 sample sites show relatively finer bed surfaces (left side of Figure 5.17). The finer surfaces are not systematically associated with any particular location or geomorphic environment within the river and may result from a variety of different factors, such as a locally high sediment supply (Dietrich et al. 1989), or local hydraulic conditions that produce particularly low values of channel competence. Alternatively, they may simply represent a local supply of relatively finer-sized material. Although the grain-size distributions are extremely variable from one bar to another (Figure 5.17), there is a slight trend of downstream fining (Figure 5.18).

Photo 15: Typical armor layer on gravel bars in Hells Canyon (note pen for scale).

Photo 16: Imbrication of gravels on gravel bars in Hells Canyon.

In addition, these imbricated materials appear to have been in place for some time. Many of these imbricated gravels are "shingled" next to one another and have worn substantial grooves into each other, which typically occurs when rocks are in contact for long periods of time under conducive weathering conditions. In this system, when these materials are periodically submerged, overlying river flows cause gravels and cobbles to shift and rock against each other and wear grooved weathering patterns. Groove depths in adjacent clasts range from approximately 13 mm to 25 mm (0.5 to 1 inch); these depths are particularly significant because the clasts tend to be predominantly dense, hard metamorphosed rocks. No relevant literature was available to help quantify this process, but best professional judgement suggests given that these bars are only inundated a few times a year during peak flows, these bars and associated bed materials have been in place for a long period of time (10s to 100s of years). In addition, some of the gravel bar surfaces have an orange varnish layer (e.g., Pine Bar) and large trees, which suggests that some of the bars have been stable for at least 1,000 years. These observations suggest that the channel bed was likely armored well before the construction of the HCC, possibly following the decrease from historical (geologic) flow regimes to current conditions (see Chapter 6 for additional discussion).

5.3.1.4. Riverbank Materials

In terms of fine-grained sediments that are adjacent to the river, shoreline erosion in some systems is a large contributor of sediments to the channel. However, in this system, riverbanks and terraces in Hells Canyon are largely stable except at a few locations. A study of shoreline erosion indicates that the Hells Canyon reach is one of the most stable reaches between the Weiser River (RM 351) and the Salmon River (RM 188; Holmstead 2001). Within Hells Canyon, erosion occurred at 60 sites encompassing a total of 6.3 km (3.9 mi) (3.1%) out of 200 km (125 mi) on both sides of the river. In addition, most sites were above the range of typical flow fluctuations that are associated with HCC hydraulic capacities (Parkinson et al. 2003). This suggests that the localized erosion is likely occurring as a result of anthropogenic disturbances (e.g., recreational foot traffic) or other hillslope processes (e.g., groundwater through-flow seepage).

A more detailed analysis was conducted by Grams and Schmidt (1999b) on Tin Shed (Photo 17) and Camp Creek, two specific riverbank terraces that have cultural significance. This study indicates that 10 to 40 meters (30 to 130 ft) of bank retreat occurred between 1955 and 1990 based on historical aerial photographs. At Tin Shed most of the terrace deposits are characterized as interbedded colluvial hillslope deposits and fluvial deposits. This terrace appears to have undergone relatively greater erosion after 1982, following the more significant erosion period between 1964 and 1982 for the adjacent sandbar (Grams and Schmidt 1999b). Because Grams and Schmidt believe that the terrace erosion process involves the bank failure of near-vertical cutbanks during or following high flows, they concluded that the loss of adjacent sandbar deposits may remove "buttress-type support from the terrace and exposes a greater height of the bank" (Grams and Schmidt 1999b). They conclude that in an unregulated system, the erosion of the bars and banks would be balanced by regular sediment deposition.

However, this conclusion essentially assumes that the HCC severely limits the amount of finer material that would be available to replenish the sandbar features, which does not appear to be the case relative to other regulation projects upstream of the HCC. The conclusion also

Photo 17: An example of riverbank erosion at Tin Shed Terrace in Hells Canyon (RM 215).

neglects all of the local sources of silt and sand sediment that continue to contribute loads to the mainstem and its geomorphic features. .

Grams and Schmidt (1999b) do acknowledge that the operations of the HCC cannot be unequivocally linked to terrace erosion (e.g., the largest peak flows since 1964 have not overtopped terrace surfaces). An example of where soil deposits appear to be aggrading is Salt Creek, where a recent soil column indicates that soil appears to be re-depositing following historical removal by placer mining operations. Given that placer mining was conducted prior to the HCC, this example supports the idea that not all soil deposits visible on beaches are remnants of larger deposits that are being washed away by riverine erosion due to the operations of the HCC. Detailed additional information regarding terrace erosion in Hells Canyon is provided in Parkinson et al. (2003).

In terms of coarser materials along the riverbanks, debris fan and hillslope materials that reach the channel are capable of abrading riverbank gravels and cobbles. In addition, these materials are undercutting bedrock ledges above the channel in localized areas. For example, near RM 208, water and sediment in the river erodes and undercuts the base of rocks dipping toward the river (Appendix E). Eventually, these rocks will undergo slope failure into the river as a sudden event, which further fractures and breaks the falling bedrock into large blocks and fragments. This processes commonly forms a rock-fragment breccia surface on the eroding face, which the river can then more easily erode and transport.

Although the riverbank terraces and bedrock undercuts are not ubiquitous, they likely contribute some sediment over longer timescales. This suggests that compared to local tributaries and adjacent hillslopes, limited volumes of sediment are supplied from the riverbanks in Hells Canyon. Additional discussion regarding the erosion and transport potential within the mainstem is provided in Section 5.3.2.

5.3.1.5. Visual and XRD Analysis of Bed Materials

Over the last several years, sediment samples from bed materials (including spawning gravels), sandbars, and terraces have been collected for various purposes. Upstream of the HCC seven (7) bed samples were collected from the mouths of each of the five major upstream tributaries to the Snake River (Owyhee River, Boise River, Payette River, Malheur River, and Weiser River), 16 bed samples were collected from the upper mainstem reach, and 12 surface grab samples and three (3) deep-core samples were collected from Brownlee Reservoir. Downstream of the HCC in Hells Canyon between RM 247 and RM 152, 23 bed samples were collected from the mainstem, three (3) samples were collected from the mouths of local tributaries that discharge to the mainstem (Deep Creek [RM 247.4], Battle Creek [RM 242.2], and Granite Creek [RM 239.6]), nine (9) core samples were collected from three major sandbars (Pine Bar [RM 227.5], Fish Trap Bar [RM 216.4], and Tin Shed Bar [RM 215.6]) and adjacent terraces, one (1) sample was collected from a sandbar in the Salmon River just upstream of the confluence with the Snake River, and two (2) samples were collected from a Bonneville Flood relict terrace (Big Bar [RM 223.7]).

Although these samples were collected during numerous field efforts for a variety of purposes, they are used in this report to aid in the understanding of potential effects of the HCC on sediments in Hells Canyon. A subset of these samples was visually analyzed for lithology, mineralogy, and degree of rounding. For bed sediments, samples were divided into a coarse-grained fraction (retained on a U.S. Standard Mesh #4 sieve, 4.76 mm [0.18 inch]) and a fine-grained fraction (passing through a U.S. Standard Mesh #4 sieve, 4.76 mm [0.18 inch]). (It is important to note that despite this division, all of these materials represent subsurface materials collected from beneath an extensive armor layer.) Reservoir, sandbar, and terrace materials consisted of fine-grained materials that were smaller than 4.76 mm (0.18 inch). To provide a better understanding of the provenance and characteristics of sediments above and below the HCC, IPC submitted a representative subset of the fine-grained samples (including bed materials, reservoir samples, and sandbar samples) for semi-quantitative X-ray diffraction mineralogy (XRD) and grain-size analysis. A summary of these analyses is provided below and a complete report is included as Appendix F.

5.3.1.5.1. Visual Analysis

The visual analysis of coarse-grained bed materials (coarse gravels to large cobbles ranging between 25.4 and 200 mm [1 and 8 inches]) suggests that most of the coarse surface bed sediments at the 23 mainstem and three tributary sites below the HCC are locally derived from the Columbia River Basalts, Seven Devils Group of meta-sediments, and the diorite intrusives exposed in local tributaries to Hells Canyon. Two of the mainstem sites included in this analysis represent spawning gravel sites. As discussed previously in the draft and final E.1-1 and E.1-2 technical reports, these materials reflect pre-impoundment conditions because they are locked beneath a well-armored layer that appears to have been armored since well before the construction of the HCC. Visual analysis of fine-grained subsurface bed sediments (less than 4.76 mm [0.18 inch] in diameter) initially suggested that these sediments were also locally derived (as reported in the draft of this report). However, additional XRD analyses described below have changed this conclusion.

5.3.1.5.2. XRD Analysis

XRD data are used above the HCC to determine the relative contribution of upstream tributaries to the mainstem sediment load coming into Brownlee Reservoir. These data are compared to XRD data collected on core samples from Brownlee Reservoir to help characterize the nature and sources of upstream sediment being trapped within the HCC. Bed material samples collected from Hells Canyon below the HCC reflect pre-impoundment conditions because they are subsurface materials collected from beneath an extensive armor layer that appears to have been in place well before the HCC was constructed.

The XRD data indicate that three major minerals (quartz, plagioclase, and potassium feldspar [k-spar]) have sufficient bedrock associations, distribution, and variability to be useful in estimating the provenance of materials throughout the study area. With the exception of Brownlee Reservoir samples, these three minerals typically comprise 90% plus of the total mineralogy of each sample. From a mineralogical perspective, sediments from the upstream Snake River Basin are distinct from local sediments within Hells Canyon on their respective quartz, plagioclase, and k-spar composition (Figure 5-19). In general, sediments upstream of the HCC have high quartz and k-spar levels, but relatively low plagioclase. K-spar is essentially limited to the rocks of the Idaho Batholith and provides a unique chemical signature in the fine-grained bed sediments in the Boise and Payette rivers (the two tributaries with the largest input of discharge to the mainstem upstream of Brownlee Reservoir). In contrast, sediments derived from below the HCC generally have high plagioclase, but relatively low quartz and particularly low k-spar.

Upstream of the HCC, XRD indicate an Idaho Batholith sediment signature at the mouths of the upstream tributaries and within the upstream mainstem reach. Because the Boise and Payette watersheds both have regulation projects that cut off their sediment-producing headwater areas (primarily Idaho Batholith sediments), the Idaho Batholith sediment signature represents either sediment loads that reached the mainstem Snake River prior to these regulation projects, or sediments deposited downstream of the upstream tributary reservoirs (e.g., Lucky Peak and Black Canyon) that are still in transit.

Based on their mineralogical signatures, the sediments found in Brownlee Reservoir appear to be mainly from the upper mainstem channel (70%), with a relatively smaller contribution of upper tributary sediments (e.g., from the Boise and Payette and Weiser; 30%). These results are consistent with the observation that more than 87% of the sediment-producing upper watershed, including the upper tributary basins, was cut off from the mainstem Snake River above Brownlee Reservoir prior to the construction of the HCC.

Below the HCC, fine-grained subsurface bed sediments have an upstream mainstem mineralogical signature. However, the signature from local sediments (from the local tributaries or hillslopes below the HCC) indicates an increasing contribution in the downstream direction from local sources (that is, the mineralogical signature confirms that the relative contribution of local sediments increases as one moves downstream through Hells Canyon). This contribution was estimated based on a simple mixing equation using the average ratio of mainstem bed sediments above the HCC and the average ratio of tributary bed sediments below the HCC as end members. Specifically, the ratios of quartz, plagioclase (which is closely related to the Columbia River Basalts that line most of Hells Canyon), and k-spar are used to estimate that the contribution of sediment from above the HCC. Quartz and plagioclase decreases at a rate of about 1% per river mile below the HCC, while k-spar remains relatively constant with a much smaller decrease in the downstream direction (that is, the rate of k-spar decrease is considerably lower than the rate of quartz decrease).

The very small rate of k-spar decrease is an important piece of information that suggests that the local tributary and hillslope sediments dilute the upstream signature in the downriver direction (that is, as one moves downstream of the HCC, local sources have an increasing influence on bed sediments). Again, this increasing downstream contribution from local sources reflects pre-impoundment conditions because these sediments are locked beneath the armor layer; the upstream entrapment of sediments within HCC does not influence this trend. If this were not the case and the entrapment of sediments within HCC was primarily responsible for diluting the upstream signature, the rate of k-spar decrease would be similar to the rate of quartz decrease.

In terms of the sandbars within Hells Canyon, sediment mineralogy and mixing equations of Pine Bar, Fish Trap Bar, and Tin Shed Bar were also used to determine the origin of the original bar-building materials, as well as the origin of more recently active areas of the bars where possible. Although XRD is a semi-quantitative method with a laboratory precision of about 2 to 3%, the averages of the three indicator minerals (quartz, plagioclase, and k-spar) are similar within each of the bars. Thus, the mineralogical signature of surface materials is consistent with the mineralogical signature of subsurface materials at each of these bars. The mineralogical signature of all three sandbars indicates upstream sediment (vs. sources from the local tributaries or hillslopes) contributed between 50% (Pine Bar) and 85% (Tin Shed Bar) of the original bar materials. This suggests that local sources of sediment do not appear to have contributed as much to the finer-grained sandbars as previously thought from visual estimates of mineralogical signatures. Stated another way, local sources only contributed between 15 and 50% of fine-grained sediments in the original depositional environment of these ancient sandbars. Based on archaeological dating of artifacts found in Tin Shed Bar sediments, this bar is believed to be more than 2,500 years old. Because of the similarity in mineralogy and grain sizes between the three bars, Pine Bar and Fish Trap Bar are likely as old as Tin Shed Bar.

5.3.1.5.3. Grain-Size Analysis

To complement available grain-size data that were available for upstream samples (upper tributary bed, upstream mainstem bed, and Brownlee Reservoir sediment) and downstream bed samples, grain size analysis was conducted on the core samples collected from Pine Bar, Fish Trap Bar, Tin Shed Bar, and the Salmon River sandbar. For the sandbars, grain sizes were compared as follows: (1) The particle size distribution (PSD) within a bar between the sandbar portion (within the level of river fluctuations) versus the terrace portion (above the level of river fluctuations), 2) the PSD versus depth in individual cores, 3) the PSD trends in a downstream direction between bars, and 4) the PSD between active portions of Salmon River and Snake River sandbars.

The comparison between the sandbar areas and adjacent terraces indicates that the median grain sizes of terrace materials are generally smaller than that of adjacent sandbar material. Within each sandbar, median grain sizes ranged between medium and coarse sands and were relatively consistent throughout the core (i.e., there were no substantive differences between the surface and subsurface materials in vertical profile). The d_{50} values in Pine Bar range from 0.30 to 0.65 mm (0.01 to 0.02 inch) (medium to coarse sand), the d_{50} values in Fish Trap Bar range from 0.24 to 0.40 mm (0.009 to 0.02 inch) (medium sands), and the d_{50} values in Tin Shed Bar range from 0.27 to 0.31 mm (0.01 to 0.01 inch) (medium sands). These median grain sizes also show that between each sandbar, the sediments are relatively consistent despite a large variability in topography, hydraulics, and other factors that influence the deposition of bars. Finally, the median grain sizes between the Snake River sandbars and the limited Salmon River sample appear to be very similar, particularly to the Fish Trap sandbar sediments.

In sharp contrast to the sandbar samples obtained in Hells Canyon, sediments sampled in Brownlee Reservoir are almost exclusively silt/clay, very fine sand, and fine sand sizes. Since the sandbars (including material deeply buried in the bars and not subject to “winnowing”) are not made up of material similar in size to the material trapped in Brownlee Reservoir, Brownlee Reservoir does not appear to have trapped a significant portion of material that would have prevented the erosion of the bars.

5.3.2. Mainstem Sediment Dynamics

Once sediment reaches the Hells Canyon mainstem channel, these sediments are stored or delivered through the system downstream toward Lewiston, Washington. In general, the coarser material from debris flows and landslides is deposited within the channel. For example, Wild Sheep, Granite Creek, and Rush Creek rapids were formed when coarse debris flow sediments from these tributaries was too large to be transported and thus remained in place, creating a stable, armored channel (Parkinson et al. 2003).

Some fine material (primarily sand) has also been deposited in the canyon, primarily in isolated eddies and near-shore environments. Except for the coarsest material, these deposits may be transitory. Some of this finer sediment occasionally erodes and flows downstream, while additional sediment from other sources accumulates. Following a discussion of sediment supplies and transport capacity in the mainstem (Section 5.3.2.1.), more detailed analysis of sandbars (Section 5.3.2.2.) and spawning sites (Section 5.3.2.3.) is presented.

5.3.2.1. Sediment Supplies and Transport Capacity

In terms of the local sediment supply to the canyon, it appears that sediments are being produced and shed at a relatively faster rate in the upper segment of Hells Canyon (between RM 247 and RM 220) than in the lower segment (between RM 220 and RM 188). This is true from both a quantitative perspective (estimated sediment yields in Granite and Sheep Creeks are significantly higher than other tributaries further downstream) and qualitative standpoint (hillslope varnishes and landslides are younger and slopes are steeper upstream of RM 220; Appendix E). A direct benefit of this differential rate of sediment production is that many of the sand beaches and spawning areas that have been identified as areas of particular concern are located downstream of local high-producing areas.

Once sediments reach the mainstem from the tributaries and hillslopes, the issues of sediment transport capacity and deposition become critical. Downstream of the HCC, a steep river slope affects the river's sediment transport capacity. Sediment transport can be described as a power function of velocity ($Q_s = aV^b$, where Q_s is sediment transport, a is a coefficient, V is the velocity of flow, and b is an exponent). Velocity is related to the square root of slope through the commonly used Manning's equation.

Because the average slope in the Hells Canyon reach between the HCD and the Salmon River is so steep (approximately 0.002, or 3 meters [10 feet] per mi), the transport capacity is quite large. The very existence of the relatively young, steep bedrock canyon suggests that the river has been able to transport almost all of its sediment downstream since it began incising the surrounding bedrock 2 million years ago (Chapter 1). Also, the nearly vertical walls associated with most of the Bonneville Flood terraces suggests that the river has continued its rapid downcutting over the last 14,500 years. The transport capacity of the mainstem appears to have been so effective that the enormous volumes of sediment produced by local sources (a minimum of 8.6 million tons/year) have not been sufficient to preclude this downcutting. Thus, the mainstem channel has apparently been generally sediment deficient with respect to the available supply of sediment over geologic timescales.

This sediment supply deficit is consistent with other mountain drainage basins that are characterized by bedrock and cascade reach-level morphologies (Montgomery and Buffington 1997), similar to Hells Canyon. In addition, it is important to note that this sediment supply deficit appears to be almost entirely connected to the large transport capacity associated with the geologic history of the canyon. For example, the lack of an active floodplain means that peak flows above bankfull levels are directly translated into greater shear stresses and larger potential sediment transport (Section 5.2.). Thus, the sediment supply deficit does not appear to be a result of the lack of upstream sediment retained by the HCC.

The relative supply of sediment can further be assessed by investigating the degree of armoring. Well-armored surfaces typically result from winnowing of the fine grain sizes and may indicate a low sediment supply relative to channel transport capacity (Dietrich et al. 1989). Comparison of surface and subsurface grain-size distributions sampled on bars in the upper and middle valley segments (Parkinson et al. 2003) demonstrates a high degree of surface armoring, but no downstream trends in armoring (Figure 5.20). The ratio of surface to subsurface median grain size (d_{50s}/d_{50ss}) observed in Hells Canyon is typically between 4 and 7, which is considerably greater than the commonly reported values of 2-4 for well-armored gravel- and cobble-bed streams (Bathurst 1987). Because the channel bed appears to have been armored for a long period of time (Section 5.3.1.3.), this suggests that the sediment supply deficit probably has been present for a similarly long period of time.

The mainstem supply deficit conclusions are confirmed by Dietrich et al.'s (1989) dimensionless bedload transport index ($q_b{*}_D$), which is used to assess relative sediment supply at bar sites where both surface and subsurface samples were obtained. $q_b{*}_D$ is a sediment transport efficiency equation for conditions of equilibrium transport (bedload transport rate equal to the rate of sediment supply). q_b*_D is defined as

$$q_b*_D = \left(\frac{\tau_{sf} - \tau_{c50s}}{\tau_{sf} - \tau_{c50ss}}\right)^{1.5} \tag{5-8}$$

where τ_{c50s} and τ_{c50ss} are the critical shear stresses for motion of d_{50s} and d_{50ss}, respectively, and τ_{sf} is evaluated at $Q_{1.5}$. The critical shear stresses (τ_{c50s} and τ_{c50s}) are calculated from the Shields (1936) equation as

$$\tau_{c50s} = \tau*_{c50s}\,(\rho_s - \rho)\,g\,d_{50s} \tag{5-9a}$$

$$\tau_{c50ss} = \tau*_{c50s}\,(\rho_s - \rho)\,g\,d_{50ss} \tag{5-9b}$$

with $\tau*_{c50s}$ set equal to 0.03 (a lower limit of typically observed incipient motion values; Buffington and Montgomery 1997), providing a conservative estimate of particle mobility.

q_b*_D values range from 1 for poorly armored surfaces (indicative of high bedload transport rates and high sediment supplies) to 0 for well armored surfaces (indicative of low bedload transport rates and low sediment supplies). Most of the sample sites in Hells Canyon (81%) have q_b*_D values less than 0.3, indicating low sediment supplies relative to the available transport capacity (Figure 5.21). Although the q_b*_D analysis and the armoring analysis both yield similar interpretations of the sediment supply conditions in Hells Canyon, the q_b*_D analysis is preferred because it evaluates sediment supply relative to channel transport capacity (a factor that is not accounted for in the armoring analysis).

Of course, despite the large transport capacity in the mainstem, depositional features do exist within the canyon. Both coarse- and fine-grained depositional features are likely transitory because they grow or shrink in response to both long-term trends in sediment supply and transport capacity, as well as to infrequent events that can mobilize and transport substantial sediment such as the Bonneville Flood 14,500 years ago. For example, debris fans produced by smaller tributary drainage basins project out into the Snake River and rapidly become armored

Photo 18: Hells Canyon at Highrange Creek (RM 206). Note that where the channel controls result in the appropriate hydraulic conditions, eddy bars allow sandbars to occur; however, these features are opportunistic and not very common because of the limited locations for bar development.

and cause the Snake River to flow around them. Spawning sites and sandbars may form on the downriver side of these armored debris fans, as discussed in Section 5.2. (Photo 18). Previous studies of changes in bar surface area in Hells Canyon indicate a decreasing supply of fine-grained material (sands and silts) from 1964 to 1982 (Grams 1991; Grams and Schmidt 1999a), as discussed in more detail in Section 5.3.2.2. The channel-unit mapping conducted for this study (Appendix D) corroborates the conclusion that there is a low supply of fine sediment in portions of the river relative to transport capacity. In particular, the eddy bars that one expects to find in the lee of debris fans are conspicuously missing or of small extent between HCD and the Salmon River confluence (Appendix D). A significant portion of the supply of fine sediment may be stored behind the numerous dams along the Snake River system, with much of it trapped in dams upstream of the HCC. The apparent reduction in supply of fine sediment to the Hells Canyon reach of the Snake River through the mid to late 1900s may also represent the declining extent and magnitude of upstream land uses that historically produced elevated loads of fine sediment (mining, logging, etc.). Consequently, while the cause for the reduced supply of sands and silts in this section of the river is uncertain, it appears that major causes may be impoundments above the HCC and changes in land use. In contrast, below the Salmon River confluence fine sediment deposits become more extensive. The Salmon River is largely unregulated and drains the Idaho Batholith, which naturally produces large quantities of coarse sands and silts.

5.3.2.2. Sand Beach Dynamics

The alluvial sandbars below the HCC have been the subject of previous studies because of their recreational and cultural significance. These studies, as well as IPC studies, are summarized in the following sections.

5.3.2.2.1. Grams Analysis

Grams and Schmidt (1991) analyzed the distribution of sandbars between HCD (RM 247) to the Salmon River (RM 188) for the period between 1964 and 1990. As part of this analysis, Graham reviewed aerial photographs only for the post-Brownlee Reservoir construction period; no pre-Brownlee Reservoir photos were able to be used to determine the pre-HCC baseline condition. Grams (1991) concluded that the frequency and area of sandbars decreased by over 75% between 1964 and 1973 and that this period is associated with the first series of large-clear water floods released from the dams. Also, between 1964 and 1973, the study concluded that the relative amount of erosion was higher in the reaches between HCD and Pittsburg Landing (RM 215); the decreased rates of erosion following the 1964 to 1973 period are due to a decreasing availability of erodable materials “as sandbars are only eroding and not rebuilding” (Grams 1991). This study also concluded that although the dams have not significantly altered the frequency or magnitude of peak flows in Hells Canyon, only clear-water flows are released into Hells Canyon. (It is important to note that this study implicitly assumed that flows coming into the HCC contained high sand sediment loads. In fact, available sediment data suggest that only a relatively minor portion of the sand sediment load would have entered the downstream canyon in the absence of the HCC, as discussed in detail in Chapter 4.)

A follow-on study conducted by Grams and Schmidt (1999a) re-evaluated the sandbar inventory for the period between 1990 and 1998. This study concluded that there were approximately the same number of bars along the river in 1998 as there were in 1990 and 1982. However, the study also concluded that while the number of bars is no longer rapidly decreasing (as it apparently did between 1964 and 1973), the sand in the remaining bars is being eroded (Grams and Schmidt 1999a). The overall area and volume of erosion exceeded the area and volume of deposition, and the study also indicates that localized depositional areas do occur. For example, at Fish Trap Bar (RM 216), repeat topographic surveys documented both erosion 2,200 yd^2 (1,820 m^2) and deposition 1,200 yd^2 (1,000 m^2) between 1990 and 1998 (Grams and Schmidt 1999a). The authors interpret these results to mean that although the sandbars are reworked during floods, the net volume of sand is decreasing (Grams and Schmidt 1999a). In addition, “continued large floods, in the absence of sediment inputs to the system, have caused erosion of the sand resources in Hells Canyon.”

Again, in addition to neglecting the contribution of other regulation projects upstream of the HCC, this conclusion does not consider the major local sources of sediment, including the tributaries and hillslopes, that appear to provide the majority of sediment input into Hells Canyon (Section 5.3.1.). In addition, if HCC operations had caused substantial erosion of materials from the sandbar complexes, including winnowing of the finer sediment sizes, one would expect to observe relatively more coarse particle size distributions associated with the sandbars as compared to adjacent terraces. However, particle size distributions for some of the major bar complexes (e.g., Pine Bar [RM 227], Salt Creek Bar [RM 222], and Fish Trap Bar [RM 216]) show that the sandbars are comprised of similar grain sizes as associated terraces (Parkinson et al., 2003). Thus, the Grams and Schmidt (1999a) conclusions appear to oversimplify complex and dynamic processes that affect the sandbars in Hells Canyon.

5.3.2.2.2. IPC Analysis

In response to concerns raised about the beaches within Hells Canyon, IPC undertook its own evaluation to quantify changes in sandbars using two different approaches:

1) Performed a physical survey of selected bars
2) Used aerial photograph interpretation to count the total number of identifiable bars within a selected reach of river

Additional information using surface area measurements from aerial photographs and distance measurements using aerial photographs were used primarily to reinforce the quantitative results from the sandbar count and physical survey methods. The results of these analyses are presented below and the complete analyses are contained in Parkinson et al. (2003).

Physical Surveys

IPC surveyed the following four sandbars to evaluate the changes in size and shape of the bars: Pine Bar (RM 227), Salt Creek Bar (RM 222), Fish Trap Bar (RM 216; Photo 19), and China Bar (RM 192). To monitor the trend at each of the bars, IPC conducted transect surveys in 1998 and 2000 and supplemented this information with more limited survey data from 1997 and 1999.

Photo 19: Fish Trap Bar in Hells Canyon (RM 216).

In addition to surveyed transects, PSD data are available for these four sandbars. Because the #200 sieve (0.074 mm [0.003 inch]) is slightly larger than the break between very fine sand and silt (0.062 mm [0.002 inch]), particle sizes smaller than the #200 sieve were not analyzed. Even using this slightly larger size as the break point, 7% or less of any of the sandbar material is smaller than sand sizes (Figure 4.6; Parkinson et al. 2003). This finding indicates that, for the sandbar areas to be maintained, the sediment supply has to have significant material in the sand-size range. Because the sandbars are not made up of material similar in size to the material trapped in Brownlee Reservoir, IPC concludes that the sediments trapped by Brownlee Reservoir would not have prevented erosion of the bars (see also Section 5.3.1.5.3). .

It is important to note that prior to this monitoring period, the Snake River in this reach experienced the highest and second highest peak discharges on record at the USGS gage below HCD (13290450). In 1997, the daily average peak flow was the highest on record of 98,000 cfs (15-minute peak was 103,000 cfs), and during the spring of 1998 the second highest daily average flow of 93,400 cfs was recorded. Also, during the spring of 1997 and 1998, several of the Hells Canyon tributaries experienced debris flows that carried high sediment loads and resulted in large new deposits on the tributary fans (Photo 19; Parkinson et al. 2003). Several of these tributaries are located above the bars that were surveyed (Pine Bar, Salt Creek, Fish Trap, and China Bar).

In general, the survey data indicate that the bars and banks experienced both deposition and erosion between 1997 and 2000 (Parkinson et al. 2003). Although each bar experienced erosion, the banks remained stable. Fish Trap Bar and China Bar seem to be the most active of the four bars in terms of raising or lowering crest elevations. At the bars with a distinct crest (all except Salt Creek Bar), the crest elevation seems to stabilize near the 30,000-cfs water surface elevation. Historical aerial photographs from 1946, 1949, and 1964 of Pine Bar show that the bars have been a dynamic feature since well before the HCC was constructed (Parkinson et al. 2003).

In addition to river hydraulics, recreational influences at Pine Bar are affecting the bar and the bank. Erosion has occurred on the upstream end of the bar and bank (from the large hackberry tree to the upstream boundary, a bedrock outcrop), particularly near a heavily used campsite area for boaters and hikers. Boats and rafts commonly land at all of the bars because they are popular recreational sites. Two common effects associated with landing boats and rafts result from the boats on the bank and the foot traffic. One effect associated with jet boats at sandbars is the effect of their jet pumps, creating local turbulence and high velocities. This turbulence can mobilize local sand particles and cause them to be redistributed in the immediate area. Although jet boat pumps can displace material as the boats pull into or out of a location with a shallow sand bottom, this effect has not been quantified and it is not known whether the practice significantly affects sandbars.

The second effect is from foot traffic associated with the landing of the boats. This traffic includes dragging and carrying camping equipment up onto the high terrace and using restroom facilities. Each of the four sites had pit toilets located near the center of the bar on the sand terrace. In addition, Salt Creek Bar and China Bar each has a picnic table located on the terrace above the bar. For more information see Parkinson et al. (2003).

Aerial Photograph Interpretation

Sandbar counts from aerial photographic surveys from 1955, 1964, 1973, 1977, 1982, and 1997 were also performed. Overall, the number of sandbars in Hells Canyon has decreased from about 215 to about 142 or 34% between 1955 and 1997 (Parkinson et al 2003). Between the pre-HCC series (1955) and the 1964 series, (the time period that includes construction of Brownlee Dam and Oxbow Dam) the total number of sandbars actually increased from 215 to 238, or 11%. The largest decline in terms of both total numbers and percentages is between 1964 and 1973 when sandbars declined from 238 to 150, or 37%. The other period of decline was between 1977 and 1982 (20%). Overall, there have been two periods of substantial decline bracketed by two periods of increase for a net decline of 34% (the fifth period 1973 to 1977 was essentially flat).

Sandbars have not responded uniformly throughout the length of Hells Canyon. The reach of the canyon from HCD to Pine Bar has the steepest gradient within the study reach. This reach had the lowest number of bars per mile in the canyon and was the only reach that shows a decline throughout the study period (1955-1997). The Pine Bar to Pittsburg Landing reach is the next steepest reach and it shows a pattern similar to the upstream reach but with a less severe overall decline from 34 to 16 bars or 53%.

The most downstream reach from Pittsburg Landing to the Salmon River has the lowest gradient in the study area. This reach has the highest number of sandbars per mile (in all years evaluated), and is the longest reach within the study area. It shows a significantly different pattern in sandbar change than the rest of the river. Over the full study period there has been an increase from 105 bars to 109 bars or 4%. This analysis shows that there is the capacity to lose bars (1982 series) and then regain them under current sediment supply, river processes, and operations of the HCC (Parkinson et al. 2003).

In order to give a frame of reference to evaluate the changes in sandbar counts, this section presents percentage of change from 1955. Normally, changes should be measured from a baseline condition with baseline representing some equilibrium natural or pre-project condition. In this case, the earliest complete data available are from immediately pre-project and the first two data points (1955 and 1964) show a clear increase in sandbar counts. It is not apparent from available data what the pre-HCC level of sandbar frequency was in Hells Canyon. However, using 1955 as a baseline yields an overall change of 34%, which is far less than what other researchers have reported (Grams and Schmidt 1991). Using 1964 as the baseline as other researchers have done (Grams and Schmidt 1991) clearly overstates the change because there are more sandbars in the 1964 series even though there is no reason to believe that the 1964 count is more representative than the 1955 count (Parkinson et al. 2003).

It is quite likely that construction upstream (including two major earth fill dams, Brownlee and Oxbow Dams) had an effect on the number and size of sandbars in the 1964 series of photos. Sandbar counts in 1955 were also probably affected by anthropogenic impacts within Hells Canyon and upstream, although these effects would be somewhat diminished by distance because they occurred farther upstream than the construction of Brownlee and Oxbow Dams. Given the lack of any better alternative, IPC used 1955 as the comparison year but recognize that using this year may overstate reductions in sandbar counts to some unknown degree.

The weight of evidence from these analyses suggest that multiple factors likely influence the size, shape, and stability of the sandbars in Hells Canyon; the construction and operation of the HCC is only one of these factors. These factors include climate changes, anthropogenic disturbances such as mining and logging that likely introduced temporary elevated sediment loads, recreational uses, and the retention of upstream sediment (92% from other regulation projects that are upstream of the HCC). It is not clear whether the size and quantities of sandbars documented in oral, written, and photographic historical accounts existed within a regime of dynamic equilibrium within the overall system, or if they were a result of anthropogenic factors (mid 1800s through mid 1900s, as discussed in Chapter 2) and available geomorphic and channel characteristics that allowed their development (see Chapter 1).

5.3.2.3. Spawning Habitat

IPC biologists modeled 17 spawning sites for fall chinook salmon in the Snake River below HCD (Parkinson et al. 2003). These spawning sites are scattered throughout the river, with 12 sites located above the Imnaha River, one site between the Imnaha and Salmon rivers, and the remaining 4 sites below the Grande Ronde River. The riverbed material was analyzed at each spawning site to determine whether the flows experienced in this reach would mobilize the material. The results indicate that the smallest spawning gravel size of 25 mm (1 inch) is stable in all of the spawning sites for flows up to 100,000 cfs except the site at Rocky Bar/Wild Sheep. At this site, flows greater than 100,000 cfs appear to move 25-mm (1-inch) material.

To determine the flows necessary to mobilize spawning gravels at other sites, dimensionless shear stress values were computed for a flow of 150,000 cfs, which is approximately 1.5 times larger than the flow of record in the study area. Again the Rocky Bar/Wild Sheep site was the only spawning location that would indicate mobility using a critical dimensionless shear stress of 0.047. Therefore, gravels larger than 50 mm (2 inches) would not move at any of these sites either. If a lower dimensionless shear stress value of 0.03 were used, only five of the 17 sites would indicate mobility at the highest flow of record (100,000 cfs) for 25-mm (1-inch) material and none of the sites would mobilize if their median particle size was 50 mm (2 inches) or greater. Since 95% of the sites in Hells Canyon have median particle sizes of at least 50 mm (2 inches), the majority of spawning sites within the canyon are stable.

In summary, only one spawning site shows gravel movement at flows experienced in the study area, and this movement occurs only for the smallest material. For the remaining sites, the computations indicate that gravels of all sizes are stable. Since the larger gravel sizes do not move, they may shield the smaller material and prevent it from moving, which in turn enhances the stability of the spawning sites (Parkinson et al. 2003).

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6. EFFECTS OF THE HCC ON STREAM FLOWS, SEDIMENT DYNAMICS, AND CHANNEL MORPHOLOGY OF HELLS CANYON

Chapter Summary

In Hells Canyon downstream from the HCC, water and sediment are the primary drivers that control channel geomorphology. Although these factors are the same as for typical alluvial rivers, the Snake River in Hells Canyon is controlled, in large part, by its complex geologic history. Most hillslope, valley, and channel morphology features appear to be relic features associated with pre-regulation hydrologic and geologic events (i.e., prior to the HCC, as well as prior to the numerous other regulation projects in the upper Snake River Basin). Although pre-settlement and pre-regulation conditions are as not well known as current conditions, the weight of evidence suggests that Hells Canyon has been a largely static river system for at least 1,000 years, if not longer. Thus, this chapter analyzes effects that the HCC has had on stream flows, sediment dynamics, and channel morphology in Hells Canyon.

Stream flows have remained essentially unchanged for the past 1,000 years (Section 6.1.). On an annual basis, the HCC is only able to retain approximately 11% of the average inflow, which is markedly smaller than other systems of comparable size. This means that the HCC is not capable of substantially affecting monthly, annual, or peak flows in Hells Canyon, which are the flows that control the channel form.

Upstream from the HCC, the majority of bed materials (>0.063 mm [>0.002 inch]) drop out before entering Brownlee Reservoir (Section 6.2.). Approximately 96% of the sediment trapped in Brownlee Reservoir consists of fine sand, very fine sand, and silt-clay, which would have flushed through the canyon reach quickly. This material represents a small fraction of the pre-regulation sediment supplies in the basin that might have otherwise reached Hells Canyon and a minor portion of local supplies of sediment that continue to be produced in the canyon. Despite the enormous volumes of sediment produced by local sources such as tributaries and hillslopes, the transport capacity in the mainstem is so large that even in a free-flowing condition, the Hells Canyon reach was likely sediment deficient with respect to the upstream supply of sediment.

Downstream from the HCC, large-scale geologic and geomorphic processes control the overall channel morphology and these processes are not appreciably altered by the HCC (Section 6.3.). Many of the potential channel morphology responses to altered discharges and reduced sediment inputs simply cannot compete with the larger geologic controls at work in the canyon. For example, decreased sediment loads generally cause rivers to armor their beds and decrease the zone of active sediment transport. In this case, the Snake River channel bed in Hells Canyon appears to have been well-armored before the HCC was built, so the expected response is likely to be muted. Although isolated sandbars respond to the minor decrease in fine-grained upstream sediment load, it is likely that these features are fluctuating within a much larger dynamic associated with the complex geologic history of the canyon.

6.1. Influence on Stream Flows

Within the context of geologic and recent history, the HCC does not appear to have substantially altered stream flows in Hells Canyon. The hydrologic regime responsible for carving Hells Canyon is estimated to have had stream flows up to 10 times greater than current stream flows (within the last 1,000 years). Since then, the hydrologic regime has shifted toward current pre-regulation conditions and Hells Canyon is believed to have been relatively stable over the last 1,000 years (see Sections 1.4. and 3.2.). Although local canyon features are certainly dynamic (such as sandbars, terrace banks, tributary debris fans, and hillslope talus slopes), the basic canyon form and morphology likely have not changed significantly under relatively smaller stream flows during the last 1,000 years.

Within the last 100 years, stream flows in the basin upstream from the HCC have been altered dramatically. The 13 mainstem facilities (not including the HCC) and 35 major tributary facilities can store a combined 10.3 million acre-feet of water. The majority of these facilities (95%) were constructed between 1901 and 1958, prior to the completion of Brownlee Reservoir. These projects had already reduced peak flows, altered the annual hydrograph, and cut annual flow volumes significantly, primarily because the majority of these facilities were constructed for irrigation purposes.

The construction of the HCC increased the total active storage capacity of the basin by 1.48 million acre-feet (13%). On an annual basis, the HCC is only able to retain approximately 11% of the average inflow, which is markedly smaller than other systems of comparable size. This means that the HCC is not capable of substantially affecting monthly, annual, or peak flows. Independent (Grams 1991) and internal (Parkinson et al. 2001) analyses have demonstrated that pre- and post-HCC peak flows are nearly identical (4% difference).

For gravel-bed rivers, the channel-forming flows (i.e., bedload-carrying flows) are typically bankfull, which is comparable to a 1.5-year recurrence interval. The value of the 1.5-year flow for the Snake River just downstream from Hells Canyon Dam is approximately 30% higher than the plant hydraulic capacity of 30,000 cfs. This suggests that the HCC power operations have had very little effect on flows that would substantially alter the downstream gravel-bed channel.

Other flow effects, including reductions in frequency, have undoubtedly altered localized, finer-grained channel features such as sandbars and terrace banks. However, the changes in these features are occurring within the context of a long-term (geological timeframe) dynamic environment. Thus, alterations to local features that have been documented within the last few decades, and have been attributed to the operation of the HCC, more likely reflect a long-term dynamic geologic environment that is independent from the HCC.

6.2. Influence on Sediment Dynamics

Because stream flows, sediment dynamics, and channel morphology are interconnected, the significantly higher paleo discharges very likely transported larger volumes and sizes of sediments throughout the study area and Hells Canyon. For example, the catastrophic Bonneville

Flood event left a significant sediment imprint on the canyon itself. Subsequent Holocene flooding events have also resulted in long-term deposition and erosional features. Following settlement in the early 1800s, anthropogenic disturbances such as mining, forest management, and grazing caused an increase in sediment supplies to the system as a whole (Chapter 2). Although these disturbances were mostly located upstream from the HCC, localized anthropogenic disturbances (most notably mining) within Hells Canyon also increased sediment supplies into the mainstem channel as compared to pre-settlement conditions.

Water regulation and storage has had the largest anthropogenic impact on sediment dynamics in the study area. Potential sediment sources to the system have been trapped by the reservoirs along the major tributaries and within the mainstem. Prior to the construction of the HCC, more than 87% of the basin drainage area was located behind other dams. The trapped sediment in the pre-HCC facilities includes not only sands (coarse suspended load) and gravels (bedload) that are deposited in the headwater portion of the reservoirs, but also silts and fine sands that settle near the dam faces.

In the mainstem upstream from the HCC, the Snake River is located in a relatively flat alluvial plain, where bed material (coarser than 0.063 mm) that reaches the mainstem is generally deposited prior to reaching the HCC. Sediment that is retained within HCC is comprised primarily of silt-clay, very fine sand, and fine sand sediments (96%). In the absence of the HCC, between 1.47 and 2.78 million tons/year of sediment, most of which are materials finer than 0.25 mm, would likely have passed through the reservoir reach into Hells Canyon. These finer sediments do not represent the typical sizes of sediments that comprise locally important features such as sandbars and spawning sites and that control channel morphology in the canyon.

Finally, historical upstream land uses (most notably mining and logging) very likely increased loads of fine sediment that passed through the reservoir reach into Hells Canyon. However, since 1901, other regulation facilities cut off 92% of the available sediment load. Thus, the fine and coarse sediments that have been retained upstream as a result of the HCC are only a small fraction of the pre-regulation sediment supplies in the basin that might have otherwise reached Hells Canyon. Although the HCC has cut off another 8% of sediments from Hells Canyon, these particles are largely silt-clay sizes that would have likely flushed through the canyon reach quickly.

Downstream from the HCC in Hells Canyon, local tributaries and adjacent hillslopes provide the majority of sediment to the Snake River. Over shorter timeframes (tens to hundreds of years), local tributaries and hillslopes contribute 8.6 million tons of material larger than 0.063 mm annually (additional fine-grained suspended load material was not quantified). The estimates are considered conservative because they are based on discharges up to a 1% recurrence interval, which do not include extreme events (such as 500- or 1000-year return events) with additional transport capacity.

Additional sediments are trapped in storage within the tributaries primarily because of the geologic development and structural controls of the canyon, both of which are independent from the operation of the HCC. In this sense, these tributaries produce a huge supply of sediment (that is, are not sediment starved), but only a proportion of sediment reaches the mainstem over the timeframe of tens to hundreds of years. Over longer timeframes, the remaining sediment has been mobilized from storage into the mainstem channel, as evidenced by the presence of relict tributary debris fans throughout the reach. Again, these episodic processes are not hindered by, and operate independently from, the upstream HCC.

Similar to the local tributaries, the short-term estimate of hillslope sediment yields is a conservative estimate that does not account for longer term, episodic and catastrophic events. Analysis of slope classes and varnish confirms that a significant volume of hillslope materials have been historically (geologic timeframe) shed directly into the mainstem channel. Indeed, the depth and steepness of the current canyon form suggests that this material has been eroding steadily during the last 2 to 6 million years at a rapid rate.

Compared to local tributaries and hillslopes, riverbed materials and riverbank (shoreline) materials do not contribute as much sediment to the Snake River. Incipient motion results for the mainstem indicate that isolated pockets of bed material move under the flows currently experienced in this reach, but the majority of the bed appears to be stable (Parkinson et al. 2003). Inspection of the gravel bars that are exposed during typical flows indicates that these bed materials are strongly imbricated and well armored.

The armoring process in gravel-bed rivers consists of a progressive coarsening of the bed layer as fine particles are removed and winnowed away. The channel bed typically becomes resistant to scour as the coarse materials remain, although the mechanisms that control the fining process continue to be debated. The removal of fine particles and development of a well-armored surface can be due either to a decrease in transport capacity from decreasing stream flows, or to a decrease in sediment supplies (as is typical below most reservoirs). In Hells Canyon, the development of an armored layer as stream flows markedly decreased (from glacial time to about 1,000 years ago) is consistent with observations that bed materials appear to have been in place for some time. Weathered grooved surfaces and varnish coating suggest that the gravel bars and associated bed material have been mostly stable for tens to hundreds of years, if not longer.

If the armoring process in the canyon resulted from the HCC, then the subsurface channel bed materials that were deposited and armored prior to the HCC would reflect upstream sources, which is not the case. In fact, fine-grained substance materials appear to be from upper tributary sources that were cut off by other regulation projects. In addition, if the HCC were responsible for the armoring observed in Hells Canyon, then this process would be progressing in a downstream direction from immediately below Hells Canyon Dam and the grain size distribution would be expected to become finer in the downstream direction (Williams and Wolman 1984). In fact, the channel bed is well armored from just below Hells Canyon Dam to below the Salmon River, with no clear downstream trend in either armoring ratios or finer grain sizes (Figure 5-20). Thus, it does not appear that the HCC has caused more than 97 m (60 miles) of riverbed to armor in approximately 30 years.

It is more likely that the channel bed was armored following the decrease from historical (geologic) flow regimes to current conditions. As the other upstream regulation projects decreased peak flows and retained most of the previously available sediment, the bed armor layer in Hells Canyon may have become more armored. However, this process appears to be largely independent from the HCC because Brownlee Reservoir has only retained an estimated 8% of the upstream sediment load that was previously available to Hells Canyon and a minor portion of the volume of sediment that continues to be available from local sources.

In terms of the local sediment supply to the canyon, it also appears that sediments are being produced and shed at a relatively faster rate in the upper segment of Hells Canyon (between RM 247 and RM 220) than in the lower segment (between RM 220 and RM 188). This is true from both a quantitative perspective (estimated sediment yields in Granite and Sheep Creeks are significantly higher than other tributaries further downstream) and qualitative standpoint (hillslope varnishes are younger and slopes are steeper upstream from RM 220). A direct benefit of this differential rate of sediment production is that many of the sand beaches and spawning areas that have been identified as areas of particular concern are located downstream from local high-producing areas.

Once sediments reach the mainstem from the tributaries and hillslopes, the issue of relative sediment transport capacity and sediment supply becomes critical. Downstream from the HCC, a steep river slope affects the river's sediment transport capacity. Compared to the upstream reach near Weiser, sediment transport capacity can be compared by taking the ratio of sediment transport in the canyon over that in the upstream reach (i.e., this simplifies to the ratio of the slopes to the $b/2$ power; see Section 5.3.2.1.). Given that the ratio of slopes above and below Hells Canyon is approximately 5, and that b is 3 to 4, the sediment transport capacity through the canyon is about 11 to 25 times larger than the sediment transport capacity for the reach upstream. This result means that, in a free-flowing condition, the Hells Canyon reach was likely generally sediment deficient with respect to the upstream supply of sediment. In other words, the sediment that was geologically and historically produced prior to regulation by the upstream watershed and passed through the HCC was quite easily transported through Hells Canyon, although some of the coarser sediment and a relatively small percentage of finer sediment is deposited in local areas.

Relationships among several key variables provide additional understanding of sediment and geomorphic issues. These relationships were developed by Lane (1955) and others, as described in Simons and Sentürk (1992). In its simplest form, the product of flow (Q) and slope (S) is proportional to the product of sediment transport (Q_s) and median sediment particle size (d_{50}):

$$QS \sim Q_s d_{50} \tag{6-1}$$

Over the long term, the flow upstream of the study area is essentially the same as the flow through the study area, except for some general increase resulting from local tributary inflow. However, the slope is significantly steeper in the Hells Canyon reach. Thus, either sediment transport increases or the median particle size increases in the Hells Canyon reach, or both. This relationship is consistent with the observed transition from the flatter Weiser reach with generally smaller bed material to the Hells Canyon reach that is dominated by significantly coarser material. While some increase in the magnitude of sediment transport is possible because of the steeper slope in the Hells Canyon reach, the actual sediment transport may not meet its capacity because the transport capacity is so much larger than the sediment supply, a deficit that is almost entirely independent from the operation of the HCC.

Through the process of hydraulic sorting, the finer sizes of sediment that reach the mainstem and are exposed to the flow are readily transported downstream or deposited in local features such as sandbars where hydraulic conditions allow, while the coarser sizes are left behind in the channel bed. Because of the lack of an active floodplain throughout the canyon, the sandbars and beaches can best be described as opportunistic features. Over the timeframe of decades, these features appear to be generally eroding and/or undergoing localized deposition and erosion. Over longer time periods (hundreds to thousands of years), these features likely vary within a dynamic range that cannot be adequately characterized within the short-term 40-year timeframe since the HCC was completed.

IPC recognizes that sandbars are important features in Hells Canyon. IPC initially used available information to address an essential question: *If the sediments trapped within the HCC (both from upstream sources and from within the reservoir reach itself) previously contributed to the sandbars in Hells Canyon, what changes in the sandbars would we expect to see as a result of the emplacement of the HCC*? Fundamentally, if the sediments traveling through the HCC reach prior to impoundment were a significant component of the *sandbar*s in Hells Canyon, the trapping of these sediments should result in a different mineralogical composition and grain-size distribution of the sandbars at depth. In other words, if the sandbar materials were composed originally of predominantly upstream sources (vs. sources from the local tributaries or hillslopes), then the cutoff of this material by the HCC should result in a surface layer from local sources that have a distinctly different mineralogical and size composition. Similarly, if the sandbar materials were originally composed of predominantly upstream sources, then after the cutoff of finer material (found in HCC) the fine materials remaining in the sandbars would be "winnowed" away and the current surface material in the sandbars would be coarser than at depth.

The results of the sediment analysis indicate that neither of these cases exist. From a mineralogical perspective, sediments within each individual sandbar are similar in composition throughout the vertical profile (there is minor variability between sandbars). Although XRD is a semi-quantitative method with a laboratory precision of about 2 to 3%, the averages of the three indicator minerals (quartz, plagioclase, and k-spar) are similar for each of the bars. Thus, the mineralogical signature of surface materials is similar to the mineralogical signature of subsurface materials at these bars. Similarly, the grain sizes of the bars are similar throughout each vertical profile and are also similar between bars. For the active portions of the bars (not those collected from adjacent terraces), the d_{50} values in Pine Bar range from 0.30 to 0.65 mm (0.01 to 0.02 inch) (medium to coarse sand), the d_{50} values in Fish Trap Bar range from 0.24 to 0.40 mm (0.009 to 0.02 inch) (medium sands), and the d_{50} values in Tin Shed Bar range from 0.27 to 0.31 mm (0.01 to 0.01 inch) (medium sands).

Thus, the sandbars are very similar in grain size and mineralogical composition and the expected responses to upstream sediment entrapment by the HCC do not appear to be evident. These data were then explored further to better understand the original depositional environment, as well as the subsequent reworking and erosion of these sandbars. The goal of this analysis was to determine what role the HCC has had, and may continue to have, in the downstream sandbars. This is particularly relevant within the context of the Grams and Schmidt (1991, 1999a) conclusions that the marked decrease in the number and volume of sandbars since the HCC was constructed is due to erosional clear-water floods caused by the HCC.

The XRD and grain size data together suggest that the original bars (including terrace areas) in Hells Canyon were probably deposited during either a single event or several events under hydrologically similar conditions. Based on archaeological dating of artifacts found in Tin Shed sediments, this bar is believed to be more than 2,500 years old. Because of the similarity in mineralogy and grain sizes between the three bars, Pine Bar and Fish Trap are likely as old as Tin Shed. The mineralogical signature of all three sandbars indicates upstream sediment (vs. sources from the local tributaries or hillslopes) contributed between 50% (Pine Bar) and 85% (Tin Shed Bar) of the original bar materials. This suggests that, unlike coarser bed materials within Hells Canyon, local sources of sediment do not appear to have contributed as much to the finer-grained sandbars as previously thought from visual estimates of mineralogical signatures. Stated another way, local sources only contributed between 15 and 50% of fine-grained sediments in the original depositional environment of these ancient sandbars.

Since the sandbars were originally deposited, these fine-grained sediments undoubtedly have been reworked through aggradation and erosion. Although the short-term load following and peak flow processes at work in the sandbars in Hells Canyon are described elsewhere, research from the Grand Canyon (Andrews et al. 1999) provides another example of where short-term deposition and erosion cycles of sandbars exemplifies their dynamic nature.

As an example within Hells Canyon, a core collected from an active area of Pine Bar indicates a similar mineral composition to that of nearby fine-grained subsurface bed materials (both with upstream mineralogical signatures). This suggests that upstream materials were deposited under pre-impoundment conditions downstream from Hells Canyon Dam. These materials were probably stored either in the channel bed or upstream sandbars and are still in transit at least as high as Pine Bar (RM 227). Of even more importance, the grain sizes from the core representing the active portion of Pine Bar are coarser-grained than those currently being deposited in Brownlee Reservoir. This appears to be another piece of evidence suggesting that these active bar sediments originally entered Hells Canyon from upstream prior to the regulation of the upper tributaries (e.g., Boise and Payette rivers).

Because the upstream sediment appears to have a large influence on the sandbars, it is important to understand the relative effects of the HCC in cutting off these sediments. Based on their mineralogical signatures, the sediments found in Brownlee Reservoir appear to be mainly from the upper mainstem channel (70%), with a relatively smaller contribution of upper tributary sediments (e.g., from the Boise and Payette and Weiser; 30%). These results are consistent with the observation that more than 87% of the sediment-producing upper watershed, including the upper tributary basins, was cut off from the mainstem Snake River above Brownlee Reservoir prior to the construction of the HCC.

Grain-size data also help further the understanding of how these bars were originally formed and how they are subsequently being reworked. The sediments sampled in Brownlee Reservoir are almost exclusively silt/clay, very fine sand, and fine sand sizes. Samples taken from sandbars in Hells Canyon show a very different PSD with a wide range of sand sizes that are better characterized as medium and coarse sands. Since the sandbars (including material deeply buried in the bars and not subject to “winnowing”) are not made up of material similar in size to the material trapped in Brownlee Reservoir, it is difficult to believe that Brownlee Reservoir has trapped a significant portion of material that would have prevented the erosion of the bars.

To summarize, the XRD and grain size data suggest that sandbars in Hells Canyon were formed primarily from upstream materials that were deposited within the canyon more than 2,500 years ago. Since the original depositional event(s), the watershed has not experienced flood conditions that changed the original depositional signature. XRD analysis suggests that the Idaho Batholith was an important source of material for the original sandbars formed in Hells Canyon. This source was cut off by water storage projects prior to and upstream from the HCC. While the HCC has cut off supplies of sediment from Hells Canyon, grain-size data suggest that Brownlee Reservoir has not trapped sediments in the size range necessary to maintain sandbars in the condition observed prior to construction of the HCC. It is unclear what effect the HCC has had on accelerating the erosion of sandbars downstream from HCD within long-term and short-term dynamic cycles of erosion and degradation associated particularly with canyon environments.

6.3. Influence on Channel Morphology

Channel morphology represents complex interactions between basin hydrology, sediment supply, and imposed external factors (for example, geologic structures, riparian vegetation, wood debris, and others). Controls on channel morphology can be represented as a nested heirarchy of physical factors (Figure 6.1). At the top tier of this heirarchy, process drivers (geology, climate, fire, and land use) impose specific watershed conditions on the fluvial system (topography, discharge, sediment supply, and vegetation). Watershed conditions, in turn, structure channel characteristics (grain size, width, depth, bed slope, channel pattern, and bed forms). Mutual adjustment of channel characteristics for different combinations of imposed watershed conditions gives rise to different reach-scale channel types or morphologies (Buffington et al. in press b).

Dams affect the primary tier of this nested heirarchy and can potentially influence channel morphology by altering both hydrologic and sediment regimes. Following a general review of the current morphology of the canyon, the contribution from each of these factors to potential changes in channel morphology is examined.

6.3.1. Current Morphology

Downstream from the HCC, large-scale geologic and geomorphic processes appear to control the overall channel morphology and these processes are not appreciably altered by the HCC. Much of the river morphology is forced by large-scale controls that significantly reduce the range of fluvial processes and types of channel adjustments that are found in other alluvial rivers of comparable size. The river valley is predominantly confined by hillslopes (54%) and cobble-boulder fans (18%), with a limited area occupied by alluvial bar complexes (18 to 20%) and terraces (8 to 9%). These data are consistent with a channel confinement ratio approximating 1.0 between the bankfull channel width and the valley width, which is an extreme and rare value for a higher-order channel such as the Snake River. The canyon downstream from the HCC also has a relatively steep slope that systematically decreases in the downstream direction from 0.002 (3.17 m [10.4 ft] per mile) near Hells Canyon Dam to 0.0007 (1.3 m [3.7 ft] per mile) downstream from the Salmon River. The combination of steep slopes and valley confinement, which largely precludes the development of a hydraulic “relief valve” during peak events, means that higher stream flows are directly translated into greater shear stresses and higher potential sediment transport.

Hydraulic geometry elements (average velocity, cross-sectional area, top width, and depth) tend to follow expected trends. The average flow width changes faster than average depth in the downstream direction (as contributing drainage area and discharge increases), while the average velocity changes the least. Predicted values of bankfull flow width and area are relatively constant above the Salmon River and then rapidly rise downstream of the tributary inputs from the Salmon and Imnaha rivers. These data show the negligible effect of smaller local tributaries on channel width and flow area between the HCC and the Salmon River, as well as the stronger hydrologic control of the major tributaries on channel dimensions.

Mapping of units within and near the channel indicates that most of the canyon features (bedrock and debris fans) control the channel morphology (rapids and pools). This is markedly different from most alluvial low-land rivers and is more typical of first-order headwater systems. Specifically, frequent and large debris fans are present all along the channel and are formed by several geomorphic process: 1) channel tributary floods and debris flows, 2) debris flows and snow avalanches on hillslopes and unchanneled valleys, 3) landslides, and 4) talus piles. The majority of these fans influence the channel morphology by obstructing and constricting stream flows. Scour pools are sometimes present along the margins of the fan where the flow has been constricted, and boulder rapids are often located downstream from these fans. Throughout the canyon, typical forced pool-riffle morphology is evident, with the frequency of pools typically occurring within a 4 to 9 channel-width spacing. The vast majority of pools (92%) are forced by external controls (that is, are not self-formed) and most of these forced pools are controlled either by bedrock or debris fans. Although sandbars should be present within the lees of the debris fans where eddies create backwater currents, such sandbars are conspicuously absent or are relatively small in extent. The reasons for this are unclear, although the lack of upstream sediment that has been primarily cut off from the canyon by other regulation projects may be a contributing factor. However, within a geologic context, these features are certainly as dynamic as the other sandbars that are not related to debris fans and the influence of the HCC within this dynamic is unclear.

6.3.2. Potential Channel Adjustments

All of these factors strongly suggest that the non-alluvial character and strong geologic controls imposed on the Hells Canyon reach limit the range of potential channel adjustments to a given perturbation, including the construction of the HCC. A typical alluvial channel can exhibit a wide variety of potential responses to disturbances that alter stream flows and sediment supplies. These responses include changes in width, depth, velocity, slope, roughness, scour depth, sediment size, and storage (Leopold and Maddock 1953; Montgomery and Buffington 1998). The direction and magnitude of these changes are indicated by relationships developed by Lane (1955) and Schumm (1977). In general, these relationships indicate that hydraulic discharge (Q) is proportional to width (W), depth (D), bedload sediment transport (Q_b), scour depth (D_s), median grain size (d_{50}), and roughness (n), and is inversely proportional to the volume of sediment stored (S_s), and slope (S):

$$Q \propto (W \times D \times Q_b \times D_s \times d_{50} \times n) / (S_s \times S) \qquad (6\text{-}2)$$

In contrast, sediment discharge (Q_s) is proportional to width (W), bedload sediment transport (Q_b), scour depth (D_s), the volume of sediment stored (S_s), and slope (S) and is inversely proportional to depth (D), median grain size (d_{50}), and roughness (n).

$$Q_s \quad \propto \quad (W \times Q_b \times D_s \times S_s \times S) / (D \times d_{50} \times n) \qquad (6\text{-}3)$$

Under typical conditions, these degrees of freedom allow a river to adjust as necessary to changing sediment and stream flow conditions. However, the Hells Canyon reach of the Snake River is not a typical system because so much of its channel morphology is governed by long-term geologic, structural, and geomorphic controls.

On a qualitative level, potential changes in mainstem Snake River channel in Hells Canyon are similar to those of a bedrock channel because of the external geologic controls. Bedrock channels are typically steep, confined systems that exhibit little alluvial bed material and a high sediment transport capacity relative to sediment supply in lower reaches of a watershed (Montgomery and Buffington 1997). For bedrock channels, it is unlikely that any of the potential channel responses will occur if sediment or stream flows are altered (Montgomery and Buffington 1997). This is also consistent with a low sensitivity to disturbance (defined as increases in stream flow magnitude and timing and/or sediment increases) and only a fair recovery potential in the Rosgen (1994) interpretation for F1-type streams.

Although the Hells Canyon reach does have a pool-riffle morphology, this is largely a forced condition that does not necessarily mimic the potential response of a self-formed pool-riffle morphology. However, to be conservative and to recognize that there are certainly local features, such as the bar complexes, that likely do respond to changes in sediment supply and discharge, this same qualitative approach can be evaluated within the context of pool-riffle channels. These types of channels are much more likely to respond to sediment and flow disturbances (Montgomery and Buffington 1997).

6.3.2.1. Discharge

Dams can alter downstream discharges by changing the frequency, magnitude, and timing of flow events compared to unregulated conditions. As discussed previously, the small storage capacity of the HCC (relative to average streamflow) precludes significant alteration of peak flows (Sections 5.1. and 6.1.). Thus, associated channel adjustments to peak discharges are likely to be negligible.

However, hydropower operation of the HCC does cause substantial short-term fluctuations of flow. During typical plant operation, discharge rates can fluctuate by up to 10,000 cfs daily 80% of the time between June 1 and September 30 (Parkinson 2001). For this analysis, an extreme flow fluctuation of 18,000 cfs was evaluated. Although these frequent and large fluctuations in discharge represent a substantial change from the natural flow regime, nonetheless, they appear to have a limited influence on channel morphology below Hells Canyon Dam.

Potential morphologic responses to hydropower operation were examined using Parker's (1990) regime diagram (Section 5.2.2.4.). A hydropeaking scenario was used in which a base level of 10,000 cfs was increased to a peak value of 28,000 cfs. Channel hydraulics were predicted from

MIKE 11 simulations for this range of flows (Section 5.2.2.). In the upper and middle valley segments (above the Salmon River confluence at RM 188), 107 sites were examined where grain size distributions were obtained from bar surfaces (Section 5.3.1.3.). As discussed previously, this approach examines the mobility of the channel cross section at each sample site assuming a bed surface similar to that of the bar. Values of dimensionless discharge (q^*) and dimensionless bedload transport rate (q_b^*) were calculated for each of these sites (Equations 5-1 through 5-5), but with the total shear stress (τ_0) replaced by the skin friction stress (τ_{sf}, as defined by Equation 5-6).

Results from this analysis show that 97% of the sites are predicted to be immobile for the above hydropeaking scenario. This result was expected given that much of the streambed is predicted to be stable even during a $Q_{1.5}$ flow (Section 5.3.1.3.), which is 41% larger than the peak flow examined in this analysis. The immobile sites will respond passively to hydropeaking, exhibiting changes in channel hydraulics (flow depth, velocity, width, and energy slope), but with no morphologic changes associated with bedload transport.

Figure 6.2 shows the potential response of the remaining 14 sites that are predicted to be mobile during the hydropower scenario. These sites are typically downstream from local tributaries, which is consistent with the peak-flow mobility predictions summarized in Section 5.3.1.3. (Parkinson et al. 2003). At each site, increased discharge accelerates bedload transport, while values of energy slope (S) decline and relative submergence ($h^*=R/d_{50}$) increases. Five of the sites are predicted to be immobile for discharges of 10,000 cfs, but exhibit rapid increases in bedload transport as discharge is ramped up to 28,000 cfs. The other nine sites are predicted to have a 13-162% increase in bedload transport rate during hydropower operation.

The mobility of these few sites indicates the potential for morphologic change. Moreover, the high frequency of hydropeaking events increases the rate of channel response compared to the historical flow regime. Over time, these frequent discharge fluctuations could potentially lead to armoring, channel degradation, and decreased bed mobility if the supply of mobile sediments at these sites was exhausted. However, despite over 30 years of hydropower operation, these supplies do not appear to be exhausted. This observation is supported by the contribution of local supplies.

It is important to note that the above analysis examines the potential geomorphic response to hydropeaking for current river conditions, but does not provide any information on past responses that may have occured since the the HCC was constructed. For example, it is possible that hydropeaking may have generally winnowed fine sediments from the streambed, partially causing the observed armoring and immobility of the river. However, field observations at numerous sites in the canyon offer no evidence of channel degradation following more than 30 years of hydropower operations associated with the HCC. In fact, particle size distributions within the main channel are generally finer than the particle size distributions found on bars above the range of operations (Parkinson et al. 2003). This suggests that the supply of sediments from local tributaries offsets potential winnowing effects.

Discharge fluctuations resulting from hydropower generation also have the potential to erode alluvial banks and fine-grained bars and beaches. Hydropeaking can erode banks and bars on alluvial margins by increasing shear stresses and by creating adverse pore pressures on the falling limb of the hydrograph. However, this sort of bank erosion is limited in Hells Canyon by extensive bedrock walls and resistant boulders that comonly armor channel margins. Those portions of the river within the Hells Canyon reach that do consist of alluvial margins are typically relict terraces and paleoflood deposits that are generally composed of erosion-resistant materials, which may explain why only 3.1% of the reach shows evidence of erosion (Section 5.3.1.4.; Holmstead 2001).

6.3.2.2. Sediment Supply

A change in sediment loads resulting from the HCC may have a much larger potential effect on channel morphology (Equation 6-3). Typically, a reduction in sediment loads might result in a corresponding decrease in width, bedload sediment transport, scour depth, sediment storage, and slope, while the depth, median grain size, and roughness factors may increase. For this system, however, the magnitude of these changes are likely to be relatively small because the HCC has only cut off 8% of the reduced upstream sediment load and an even smaller proportion of the total sediment load produced by local sources. In addition, the sediment trapped in the HCC consists of very fine particles that do not control the channel form in gravel and boulder streams. Finally, many of these potential responses are driven by changes in shear stress that likely cannot compete with the larger geologic controls at work in the canyon. For example, decreased sediment loads generally cause rivers to armor their beds and decrease the zone of active sediment transport (Dietrich et al. 1989). In this case, the bed already appears to have been well-armored before the HCC was built, so the expected response is likely to be muted.

Each of these potential changes is evaluated in more detail below.

- Width–Within the canyon, the channel width appears to be relatively stable for the first 97 m (60 miles) downstream from the HCC (RM 247 to RM 188) and then increases once the tributary inputs from the Salmon and Imnaha rivers exert more influence on the mainstem channel. Even within the first 97 m (60 miles), if anything, the river appears to be increasing its width slightly in localized areas (e.g., undercutting of bedrock walls, weathering of talus slopes). In general, the low sinuosity and channel confinement limit the responses of channel width. Also, observations suggest that lateral channel movement has been constrained for a geologically long period of time. For example, following the deposition of the large terraces associated with the Bonneville Flood, the river has cut primarily vertically, leaving almost perpendicular riverbank terrace slopes on either side of the channel.
- Bedload Sediment Transport–The channel bed is a well-armored system that was likely established under paleo flow conditions with much higher discharges and shear stresses. Under peak flow conditions, which are the same for the pre- and post-HCC periods, only localized pockets of mobile material are present. These pockets are located downstream from the tributary mouths, which suggests that the parent materials for these mobile pockets are the associated tributaries. In addition, these pockets of mobile bed materials consist of gravels, cobbles, and boulders that are not maintained by the finer particles (silt-clay and fine sands) that are trapped upstream in the HCC.
- Scour Depth–Similar to bedload sediment transport, the thickness of the active transport layer probably has not changed significantly as a result of the HCC given well-armored channel bed that likely existed prior to the HCC. Also, the fact that there was no difference in

the residual pool depth between forced and self-formed pools throughout the canyon suggests that geologic controls have a larger influence on scour depth than a relatively minor decrease in sediment loads.

- Sediment Storage–A decrease in the suspended load and bedload should produce a decrease in sediment storage as a system tries to maintain equilibrium. In Hells Canyon, sediment storage is made up primarily of localized pockets of mobile bed material and finer-grained features such as the sandbars, as well as paleo debris fan and terrace features that are not affected by the river to a large extent. Because the mobile coarse-grained bed storage elements appear to be maintained and controlled by tributary inputs, changes to these pockets are likely independent from minor upstream reductions in sediment loads attributable to the HCC. Decreases in sediment storage associated with the sandbar complexes have been documented during the last 40 years, but the material required to maintain these storage elements are more coarse (0.063–2 mm) than 73% of the material retained in the HCC (<0.063 mm). This suggests that changes to these features are not necessarily a result of the reduced upstream sediment discharge associated with the HCC.
- Slope–The current gradient in the HCC reach (0.002) is similar to both the reservoir canyon reach (0.001) and to the pre-HCC gradient (0.002; USGS 1925). Thus, potential changes to the downstream slope resulting from the HCC appear to be minimal. It appears that the slope in the canyon is driven more by active geological uplifting and fault zones, which are still likely causing downcutting. These structural elements likely exert a much larger influence on the general canyon slope than the minor reduction in upstream sediment loads attributable to the HCC.
- Depth–A direct observation of the depth of the canyon suggests that the river has rapidly downcut its channel bed during the last 2 million years (average canyon depth of 1,676 m (5,500 ft) would suggest an average rate of 0.03 inches per year). Downcutting appears to have increased following the Bonneville Flood (average elevation of Johnson Bar at 27 m (90 ft) above the current river channel suggests an average rate of 0.07 inches per year). However, when discharges started to approximate current pre-regulation conditions about 1,000 years ago, it is likely that the downcutting rate slowed because the river bed probably started to armor and the armoring process hinders downcutting. The materials below the current river bed are likely remnant alluvial flood sediments (CH2M HILL 1990), which could be eroded further if the armor layer was broken up. Significant long-term geological structural controls (e.g., uplift, faulting) likely exert a much larger influence on the ability of the channel to increase its depth than the small reduction in upstream sediment loads associated with the HCC.
- Grain Size– The surface armor layer consists of cobbles and boulders with an average median grain size (d_{50}) of 144 mm (Parkinson et al. 2001). Modeling results indicate that this bed surface is largely stable under pre- and post-HCC peak flows. Although the median grain sizes appear to decrease slightly in the downstream direction from the HCC, this may be related more closely to fact that the upstream segment of the river produces more local sediments. As these sediments are rounded and abraded, the median grain sizes would be expected to decrease in the downstream direction. The production of these sediments from local sources is independent from the HCC and governed by long-term geologic controls.

- Roughness–In terms of bed roughness in this system, the armored layer indicates that the grain resistance appears to be mostly fixed. Because the well-armored bed is believed to be caused by other sources than the HCC and the channel bed is comprised of local sources, the grain resistance in the pockets of mobile bed materials appears to be largely independent from the HCC. Other elements of roughness, including bedforms and channel bends, are largely forced by the canyon's geologic controls that limit the sinuosity and contribute to the channel confinement. Thus, roughness probably has not increased significantly since the HCC retained a relatively minor portion of the upstream sediment load.

Qualitatively, the direct effects on channel morphology downstream from the HCC appear to be minimal, primarily because the canyon is governed so strongly by long-term geologic and geomorphic structural controls that operate completely independently from the upstream HCC reservoirs. Although isolated pockets of materials (sandbars, terrace banks) are subject to respond to the decrease in fine-grained upstream sediment load, it is likely that these features are fluctuating within a much larger dynamic associated with the complex geologic history of the canyon. This is particularly probable given that most of the sediments in these features appear to be derived from local sources via processes that continue to be unconstrained by the HCC.

Quantitatively, the potential geomorphic effects of reduced sediment supply was further evaluated by determining what proportion of retained sediment would travel as bedload versus suspended load if released from the HCC. As mentioned previously, dams reduce the supply of sediment to downstream channel reaches, resulting in several potential geomorphic effects:

1. Streambed armoring and channel degradation as a result of winnowing of fine sediments (Komura 1967; Williams and Wolman 1984)

2. Increased scour depth of pools (Buffington et al. in press a)

3. Depletion of sediment storage sites (eddy bars and sand beaches, in this case; Webb et al. 1999)

The first two effects are related to reduced supplies of bedload material, while the third effect is related to diminished supplies of either bedload or suspended load material, depending on the local hydraulics and size of material that forms bars and beaches.

Consequently, results from the hydrodynamic model MIKE 11 (DHI 2000) were combined with Dietrich's (1982) settling velocity curves to calculate the largest particle size that could be carried in suspension by the 1.5-year flow ($Q_{1.5}$, a typical recurrence interval for bankfull flows in floodplain rivers; Section 5.2.2.). For these calculations, a Corey (1949) shape factor of 0.7, a Powers' (1953) roundness of 3.5, and a settling velocity equal to the $Q_{1.5}$ shear velocity was assumed. Average maximum suspendable sizes predicted for the upper, middle, and lower valley segments are 6.3, 4.5, and 3.4 mm, respectively, for the $Q_{1.5}$ event. Particle sizes larger than these values either would travel as bedload or be immobile, while smaller sizes would travel as suspended load. Sediment samples from Brownlee Reservoir demonstrate that most of the stored sediment (96%) is composed of silts and sands (86% of which is silt-clay sized sediment) of less than 2mm (Section 4.2.3.). Therefore, most of the this retained material would travel as suspended load through all three canyon valley segments if released from the HCC. Consequently, storage of this predominantly suspended-load material by the HCC is unlikely to have caused any significant downstream armoring, degradation, or pool scour (factors that are sensitive to altered supply of bedload material, rather than suspended load).

In addition to physically impounding sediment, dams may create backwater effects that can force upstream deposition of sediment miles above the reservoir, particularly the coarser particles traveling as bedload. However, as explained in detail in Section 4.1.3.3., the reach of the Snake River upstream from Brownlee Reservoir does not appear to be an active, mobile river with a high volume of coarse sediment moving through. Stable USGS cross-sections surveyed at Weiser and consistent stage-discharge relationships suggests that the river near Weiser has neither degraded nor aggraded substantially for at least 60 years. In addition, photographs of the reach between Weiser and Brownlee Reservoir suggest that substantial volumes of sand and gravel are not being transported past Weiser and into the reservoir backwater area. Thus, the distal influence of the HCC on bedload sediment supplies and the subsequent potential effects on channel morphology downstream of Hells Canyon Dam appear to be minimal.

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8. LITERATURE CITED

Abramovich R, Molnau M, Craine K. 1998. *Climates of Idaho*. Moscow (ID): University of Idaho, Cooperative Extension System, College of Agriculture.

Alpha TR, Vallier TL. 1994. Physiography of the Seven Devils Mountains and adjacent Hells Canyon of the Snake River, Idaho and Oregon. In: Vallier TL, Brooks HC, editors. *Geology of the Blue Mountains Region of Oregon, Idaho, and Washington: Stratigraphy, physiography, and mineral resources of the Blue Mountains Region*. Washington (DC): US Geological Survey, Professional Paper 1439.

Andrews ED, Johnston CE, Schmidt JC, Gonzales M. 1999. Topographic evolution of sand bars. In Webb RH, Schmidt JC, Marzolf GR, Valdez RA (eds.) The Controlled Flood in the Grand Canyon. American Geophysical Union, Monograph 110, Washington, DC, pp. 117-130.

Arrowrock Group. 1995. Idaho Power Company Hells Canyon oral history report. Boise (ID): Prepared for Idaho Power Company.

Asherin DA, Claar JJ. 1976. Inventory of riparian habitats and associated wildlife along the Columbia and Snake Rivers. Moscow (ID): College of Forestry, Wildlife, Range Sciences, University of Idaho, 556 p.

Asotin. 2001. Draft Asotin Creek Subbasin Summary. Prepared for the Northwest Power Planning Council. Asotin (WA): Asotin County Conservation District and Conservation Innovations. June 1, 2001.

Bathurst JC. 1981. Discussion of bar resistance of gravel-bed streams. *Journal of the Hydraulics Divisions*, ASCE, 107:1276-1278.

Bathurst JC. 1987. Measuring and modeling bedload transport in channels with coarse bed materials. *In* K.S. Richards (eds.) *River Channels: Environment and Process*. Oxford: Basil Blackwell, pp. 272-294.

Bennett EH. 1995. Mining, minerals, and the environment in Idaho, 1995. Moscow (ID): Idaho Geological Survey, Staff Report.

Bisson PA, Nielsen JL, Palmason RA, Grove LE. 1982. A system of naming habitat types in small streams, with examples of habitat utilization by salmonids during low streamflow. *In* Armantrout NB (eds.) Proceedings of a Symposium on Acquisition and Utilization of Aquatic Habitat Inventory Information. Bethesda (MD): Western Division of the American Fisheries Society, pp. 62-73.

Blair C, Braatne J, Simons R, Rood S, Wilson B. 2001. Effects of constructing and operating the Hells Canyon Complex on wildlife habitat. Technical Report E.3.2-44. Boise (ID): Idaho Power Company, April 2001.

Blakesley JA, Wright RG. 1988. A review of scientific research at Craters of the Moon National Monument. Moscow (ID): University of Idaho, College of Forestry, Wildlife and Range Sciences. Bulletin No. 50.

Bonneville Power Administration (BPA). 1985. Final Report Hells Canyon Environmental Investigation. U.S. Department of Energy.

Booth DB. 1990. Stream-channel incision following drainage-basin urbanization. *Water Resources Bulletin*, 26 (3), p. 407-417.

Borden JC. 2001. A comparison of sediment monitoring to sediment facies mapping in the Middle Fork Payette River, Central Idaho. Unpublished Masters' Thesis, Department of Geology. Boise (ID): Boise State University.

Buchanon K. 2001. Performance Management Alliance. Personal communication with Steve Miller, CH2M HILL, Boise, Idaho, July 18, 2001.

Buffington JM, Montgomery DE. 1997. A systematic analysis of eight decades of incipient motion studies, with special reference to gravel-bedded rivers. *Water Resources Research* 33:1993-2029.

Buffington JM, Lisle TE, Woodsmith RD, Hilton S. in press a. Controls on the size and occurrence of pools in coarse-grained forest rivers. *River Research and Applications*.

Buffington JM, Woodsmith RD, Booth DB, Montgomery DR. in press b. Fluvial processes in the Pacific Northwest and Puget Sound Rivers. In Montgomery DR, Booth DB, Bolton SB (eds.) *Puget Sound Restoration*. Seattle (WA): University of Washington Press.

Bureau of Land Management (BLM). 1986. Draft Baker resource management plan environmental impact statement. Baker (OR): Baker Resource Area, Vale District, U.S. Dept. Int. 120p.

Butler M (Ed.). 2001. Topographic integration for the Hells Canyon studies. *In*: Technical appendices for the Hells Canyon Complex Hydroelectric Project. Technical Report E.1-3. Boise (ID): Idaho Power Company.

Butler RB. 1978. A guide to understanding Idaho archaeology: the upper Snake and Salmon River country. Pocatello (ID): Idaho Museum of Natural History.

Camp VE, Hooper PR, Swanson DA, Wright TL. 1982. Columbia River Basalt in Idaho: Physical and chemical characteristics, flow distribution, and tectonic implications. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 55–75.

Carrey J, Conley C, Barton A. 1979. *Snake River of Hells Canyon.* Cambridge (ID): Backeddy Books.

CH2M HILL. 1987. Hells Canyon Dam and Powerhouse: Safety Report; FERC Project Number 1971. Boise (ID): Prepared for Idaho Power Company.

CH2M HILL. 1990. Hells Canyon Dam and Powerhouse: Stability Study; FERC Project Number 1971. Boise (ID): Prepared for Idaho Power Company.

Chaney E, Elmore E, Platts WS. 1990. Livestock grazing on western riparian areas. Produced for the U.S. Environmental Protection Agency. Northwest Resource Information Center, Inc., Eagle, Idaho.

Chapman WM. 1940. Report of a field trip to the Snake River drainage in Idaho and Eastern Oregon. Olympia (WA): Washington Department of Fisheries.

Charlton FG, Brown PM, Benson RW. 1978. The hydraulic geometry of some gravel rivers in Britain. *Report IT 180*, *Hydraulics Research Station, Wallingford, Great Britain*.

Chatters JC, Reid KC. 1999-2000. Unpublished progress reports on the archaeological inventory of Hells Canyon. Boise (ID): Prepared for Idaho Power Company.

Clark GM, Maret TR, Rupert MG, Maupin MA, Law WH, Olt DS. 1998. Water quality in the upper Snake River basin, Idaho Wyoming, 1992-95. Boise (ID): U.S. Geological Survey, Circular 1160.

Clark GM. 1994. Assessment of selected constituents in surface water of the upper Snake River Basin, Idaho and western Wyoming, water years 1975-89: U.S. Geological Survey Water-Resources Investigation Reports 93-4229, p 49.

Clark LR, Sampson RN. 1995. Forest ecosystem health in the Inland West: A science and policy reader. Washington (DC): Forest Policy Center, American Forests.

Clark R. 1995. River of the west: a chronicle of the Columbia. Picador Press, New York, NY.

Cochran BD. 1991. Geomorphological context and site formation processes. *In* K.C. Reid (ed.) *Prehistory and Paleoenvironments at Pittsburg Landing: Data Recovery and Test excavations at Six Sites in Hells Canyon National Recreation Area, West Central Idaho*. Project Report Number 15, Center for Northwest Anthropology. Pullman (WA): Washington State University, pp. 54-118.

Collier M, Webb RH, Schmidt JC. 1996. Dams and rivers: a primer on the downstream effects of dams. Tuscon (AZ): U.S. Geological Survey, Circular 1126.

Corey AT. 1949. Influence of shape on the fall velocity of sand grains. Unpublished Masters' Thesis. Ft. Collins (CO): Colorado State University.

Danish Hydraulic Institute (DHI). 2000. MIKE 11, A modeling system for rivers and channels. Copenhagen, Denmark.

DeBolt A. 1992. The ecology of *Celtis reticulata* Torr. (netleaf hackberry) in Idaho. Corvallis (OR): Oregon State University, 166 p.

Dietrich WE, Dunne T, Humphrey NF, Reid LM. 1982. Construction of sediment budgets for drainage basins, in *Sediment Budget and Routing in Forested Drainage Basins*. Portland (OR): U.S. Department of Agriculture, U.S. Forest Service, PNW Forest Experimental Station, General Technical Report PNW-141.

Dietrich WE, Kirchner JW, Ikeda H, Iseya F. 1989. Sediment supply and the development of the course surface layer in gravel-bedded rivers. *Nature* 340:215-217.

Dietrich WE. 1982. Settling velocity of natural particles. *Water Resources Research* 18:1615-1626.

Dixon MD, Wasniewski LW. 1998. Summary of landslide inventory on the westside of the Payette National Forest from the New Year 1997 storm. McCall (ID): Payette National Forest, U.S. Forest Service.

Dixon MD. 2001. Payette National Forest, U.S. Forest Service. Personal Communication with Sherrill Doran, CH2M HILL, Boise, Idaho. May 2001.

Dorn RI, Oberlander TM. 1981. Microbial Origin of Desert Varnish. *Science*, Vol. 213, pp. 1245-1247.

Einstein HA, Barbarossa NL. 1952. River channel roughness. *American Society of Civil Engineers Transactions* 117:1121-1.

Emiliani C. 1972. Quaternary paleotemperatures and duration of the high temperature intervals. *Science*, v. 178. p. 398–401.

Emmett WW. 1975. The channels and waters of the upper Salmon River area, Idaho. Washington (DC): U.S. Geological Survey, Professional Paper 870-A.

EPA. 1998. Urbanization and streams: studies of hydrologic impacts. Accessed from http://www.epa.gov/OWOW/NPS/urbanize/report.html on April 16, 2001. Washington (DC): Office of Water, Watershed Protection, March 17, 1998.

Federal Power Commission. 1959. Project 1971, Federal Power Commission hearings, December 10, 1959: Testimony of Mr. Murray and Mr. Day. Idaho Power Company files.

Ferguson RI. 1986. Hydraulics and hydraulic geometry. *Progress in Physical Geography* 10:1-31.

Fitzgerald JK. 1999. Middle Fork Payette River TMDL sediment trend monitoring results report, 1998. Seattle (WA): U.S. Environmental Protection Agency.

Fleisher M, Liu T, Broecker WS. 1999. A clue regarding the origin of rock varnish: *Geophysical Research Letters*, Vol. 26, pp. 103-106.

Flint RF. 1976. *Glacial and Quaternary Geology*. New York (NY): John Wiley and Sons.

Frissell CA, Liss WJ, Warren CE, Hurley MD. 1986. A hierarchical framework for stream habitat classification: Viewing streams in a watershed context. *Environmental Management* 10:199-214.

Grams PE, Schmidt JC. 1999a. Final Report: Sand bar erosion and deposition on the Snake River in Hells Canyon between 1990 and 1998. Department of Geography and Earth Resources. Logan (UT): Utah State University.

Grams PE, Schmidt JC. 1999b. Final Report: Sand bar and terrace erosion between 1964 and 1996 at the Tin Shed and Camp Creek cultural resource sites on the Snake River in Hells Canyon. Department of Geography and Earth Resources. Logan (UT): Utah State University.

Grams PE. 1991. Degradation of alluvial sand bars along the Snake River below Hells Canyon Dam, Hells Canyon National Recreation Area, Idaho. Senior Thesis, Department of Geology. Middlebury (Vermont): Middlebury College.

Gray M, McAfee, Jr. R, Wolf CL. 1974. Glossary of Geology: American Geological Institute, Washington (DC): 805p. plus a 52p, bibliography as an Appendix.

Hackett WR, Morgan LA. 1988. Explosive basaltic and rhyolitic volcanism of the eastern Snake River Plain. In: Link PK, Hackett WR, editors. *Guidebook to the geology of central and southern Idaho*. Moscow (ID): Idaho Geological Survey, Bulletin 27: p. 283–301.

Harts WH. 1899. Preliminary Examination of Snake River from Asotin, Wash. To Pittsburg Landing, Oregon. Letter to the House of Representatives from the Secretary of War and the Chief of Engineers Office. Washington (DC): Document no. 75. Serial set 3954.

Hey RD, Thorne CR. 1986. Stable channels with mobile gravel beds. *Journal of Hydraulic Engineering,* 112:671-687.

Holmstead GL. 2001. Shoreline Erosion in Hells Canyon. *In*: Technical appendices for the Hells Canyon Complex Hydroelectric Project. Technical Report No. E.3.2-42. Boise (ID): Idaho Power Company.

Hudson's Bay Record Society. 1979. Ogden's Snake Country Journals, 1824-1826. Publications of the Hudson's Bay Record Society, London. Kraus reprint. Nendeln, Liechtenstein.

Hydrocomp. 1990. Streamflow characteristics of the Snake River Basin. Redwood City (CA): Hydrocomp Inc. and Charles Howard and Associates. Prepared for Idaho Power Company: 85 pp.

Hyndman DW. 1989. Formation of the northern Idaho Batholith and the related mylonite of the western Idaho suture zone. In: Chamberlain VE, Breckenridge RM, Bonnichsen B, editors. *Guidebook to the geology of northern and western Idaho and surrounding area*. Moscow (ID): Idaho Geological Survey, Bulletin 28: p. 51–64.

Idaho Bureau of Mines and Geology. 1961. Alluvial mining in the Pacific Northwest. Moscow (ID): Alluvial Mining Subcommittee of the Columbia Basin Interagency Subcommittee. Idaho Bureau of Mines and Geology, Information Circular No. 9.

Idaho Bureau of Mines and Geology. 1963. Mining history of south-central Idaho. Moscow (ID): Idaho Geological Survey, Pamphlet 131.

Idaho Department of Environmental Quality [IDEQ]. 1993. Idaho Agricultural Pollution Abatement Plan. Boise (ID). Idaho Department of Environmental Quality and Idaho Department of Lands Soil Conservation Commission, August, 1993.

IDEQ. 1997. Final Middle Snake River Watershed Management Plan**,** Phase 1 TMDL Total Phosphorus. Twin Falls (ID): Division of Environmental Quality.

IDEQ. 2000. Draft subbasin assessment for Upper Boise River watersheds. Boise (ID): Idaho Department of Environmental Quality, Boise Regional Office. Public review draft.

IDEQ. 2001. Draft Snake River - Hells Canyon subbasin assessment. Boise (ID): Idaho Department of Environmental Quality. Public review draft, March 12, 2001.

Idaho Department of Water Resources [IDWR]. 1981. Geothermal investigations in Idaho: Part 11–Geological, hydrological, geochemical, and geophysical investigation of the Nampa-Caldwell and adjacent areas, southwestern Idaho. Boise (ID): Idaho Department of Water Resources, Water Information Bulletin No. 30.

IDWR. 1999. Feasibility of large-scale managed recharge of the eastern Snake Plain aquifer system. Boise (ID): Idaho Department of Water Resources, prepared in cooperation with US Bureau of Reclamation and Navigant Consulting, Inc. Retrieved from http://www.idwr.state.id.us/recharge/sections1and2.htm on March 11, 2000.

Idaho Mining Association. 2001. Golden dreams and silver linings; A history of mining in Idaho. Retrieved from http://www.idahomining.org/history.html on May 23, 2001.

Idaho Power Company [IPC]. 1997. Aerial photographs of Hells Canyon. 1:8400.

IPC. 1999. Hells Canyon Recreation Use Study. Boise (ID): Idaho Power Company.

IPC. 2000a. Hells Canyon Complex, Operations and Relicensing. Boise (ID): Idaho Power Company.

IPC. 2000b. Terrestrial Effects of Load Following, Responses to FERC Additional Request #13. C.J. Strike Project, FERC No. 2055, Prepared by Golder Associates and Idaho Power, October 2000.

Jenks MD, Bonnichsen B. 1989. Subaqueous basalt eruption into Pliocene Lake Idaho, Snake River Plain, Idaho. In: Chamberlain VE, Breckenridge RM, Bonnichsen B, editors. *Guidebook to the geology of northern and western Idaho and surrounding area*. Moscow (ID): Idaho Geological Survey, Bulletin 28: p. 17–34.

Johnejack KR, Megahan WF. 1991. Sediment transport in headwater channels in Idaho. Proceedings of the Fifth Federal Interagency Sedimentation Conference: pp. 4-155 to 4-161.

Johnson CG, Simon SA. 1987. Plant associations of the Wallowa-Snake province, Wallowa-Whitman National Forest. U.S Forest Service. PNR. R-6 ECOL-TP-225A-86. 272 p.

Johnson KM, Lewis RS, Bennett EH, Kiilsgaard TH. 1988. Cretaceous and Tertiary intrusive rocks of south-central Idaho. In: Link PK, Hackett WR, editors. *Guidebook to the geology of central and southern Idaho*. Moscow (ID): Idaho Geological Survey, Bulletin 27: p. 55–86.

Johnson WC, Dixon M. 1997. Conclusions from Snake River vegetation research. Unpublished research used in support of Waite Osterkamp's (1997) expert witness report (WO-1922 to WO-1930). Brookings (SD): South Dakota State University.

Jones JA. 1993. Soils. *In: Atlas of the Pacific Northwest* (P.L. Jackson and A.J. Kimerling, eds). Corvallis (OR): Oregon State University Press. pp:66-70.

Jones ML, Seitz HR. 1980. Sediment transport in the Snake and Clearwater Rivers in the vicinity of Lewiston, Idaho. Washington (DC): US Geological Survey, Water Resources Investigations, Open File Report 80-690.

Keller EA, Melhorn WN. 1978. Rhythmic spacing and origin of pools and riffles. *Geological Society of America Bulletin* 89:723-730.

Kellerhals R, Neill CR, Bray DI. 1972. Hydraulic and geomorphic characteristics of rivers in Alberta. *Research Council of Alberta*, *River Engineering and Surface Hydrology Report 72-1*.

Keulegan GH. 1938. Laws of turbulent flow in open channels. *Journal of Research of the National Bureau of Standards* 21:707-741.

Kimmel PG. 1982. Stratigraphy, age, and tectonic setting of the Miocene-Pliocene lacustrine sediments of the western Snake River Plain, Oregon and Idaho. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 559–578.

Kinsley D. 1998. Models of rock varnish formation constrained by high resolution transmission electron microscopy: *Sedimentology*, Vol. 45, pp. 711-725.

Kirchner JW, Finkel RC, Riebe CS, Granger DS, Clayon JL, King JG, Megahan WF. 2001. Mountain erosion over 10 yr, 10 ky, and 10 my time scales. *Geology*, 29 (7): 591–594.

Kjelstrom LC, Moffatt RL. 1981. A method of estimating flood-frequency parameters for streams in Idaho. *U.S. Geological Survey*, *Open-File Report 81-909*.

Kjelstrom LC. 1992. Simulation of water surface elevations for the Snake River in the Deer Flat National Wildlife Refuge, Idaho. Boise (ID): US Geological Survey, Water Resources Investigation Report 91-4198.

Kjelstrom LC. 1995. Streamflow gains and losses in the Snake River and groundwater budgets for the Snake River Plain, Idaho and Eastern Oregon. Washington (DC): US Geological Survey, Professional Paper 1408-C.

Knighton D. 1998. *Fluvial Forms and Processes*. London: John Wiley and Sons.

Komura S, Simmons DB. 1967. River-bed degradation below dams. Journal of the Hydraulics Division, American Society of Civil Engineers, 93:1-14.

Lane EW. 1955. Design of Stable Channels. Transactions ASCE, Vol. 120, Paper no. 2776.

Langdon JL. 2001. *The Evaluation and Conservation of Late Prehistoric House Features in Hells Canyon National Recreation: Geographic Information Systems (GIS) Applications in Archaeological Research and Management*. Unpublished Masters' Thesis, Department of Anthropology. Ellensburg (WA): Central Washington University.

Langston N. 1995. Forest Dreams, Forest Nightmares: The paradox of old growth in the inland northwest. Seattle (WA): University of Washington Press.

Leeman WP. 1982. Development of the Snake River Plain-Yellowstone Plateau province, Idaho and Wyoming: An overview and petrologic model. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 155–177.

Leopold LB, Maddock T. 1953. Hydrology for urban land uses planning: a guidebook on the hydrologic effects of urban land use. Washington (DC): US Geological Survey Circular 554.

Leopold LB, Wolman MG, Miller JP. 1964. *Fluvial processes in geomorphology*. San Francisco (CA): W.H. Freeman and Company.

Lewis GC, Fosburg MA. 1982. Distribution and character of loess and loess soils in southeastern Idaho. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 705–716.

Lipscomb SW. 1998. Hydrologic classification and estimation of basin and hydrological characteristics of subbasins in central Idaho. Denver (CO): US Geological Survey, Professional Paper 1604.

Lisle TE, Hilton S. 1999. Fine-bed material in pools of natural gravel bed channels. *Water Resources Research*: 35(4), p. 1291-1304.

Lisle TE. 1987. Using residual depths to monitor pool depths independently from discharge. USDA Forest Service, Research Note PSW-394. Albany (CA): Pacific Southwest Research Station.

Liu T, Broecker WS. 2000. How fast does rock varnish grow? *Geology* Vol. 28: pp. 183-186.

Long CE, Wiberg PL, Nowell ARM. 1993. Evaluation of von Karman's constant from integral flow parameters. *Journal of Hydraulic Engineering* 119:1182-1190.

Mabey DR. 1982. Geophysics and tectonics of the Snake River Plain. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 139–153.

Malde HE. 1968. The catastrophic late Pleistocene Bonneville Flood in the Snake River Plain, Idaho. Washington (DC): US Geological Survey, Professional Paper 596.

Meade RH. 1982. Source, sinks, and storage of river sediments in the Atlantic drainage of the United States. Journal of Geology, 90(3). Pp. 235-252. May 1982.

Meeuwig RO, Packer PE. 1976. Erosion and runoff on forest and range lands. Boise (ID): Proceedings of the 5th Workshop of the U.S./Australia Rangelands Panel, p. 105-116.

Megahan WF, Molitor DC. 1975. Erosional effects of wildlife and logging in Idaho. Journal of Irrigation and Drainage, American Society of Civil Engineers, p. 423-444.

Megahan WF. 1982. Channel sediment storage behind obstructions in forested drainage basins draining the granitic bedrock of the Idaho Batholith. Portland (OR): U.S. Department of Agriculture, U.S. Forest Service, Pacific Northwest Forest and Range Experimental Station, General Technical Report PNW-141. p. 114-121.

Meyer-Peter E, Müller R. 1948. Formulas for bed-load transport. In (eds.) *Proceedings of the 2nd Meeting of the International Association for Hydraulic Structures Research*. Delft, Netherlands: International Association for Hydraulics Research, pp. 39-64.

Miller PJB. 1975. Guidebook for the geology and scenery of the Snake River on the Idaho-Oregon border from Brownlee Dam to Hells Canyon Dam. Boise (ID): Geological Society of America Field Trip Guide No. 8.

Miller TB, Johnson FD. 1976. Ecology of riparian communities dominated by white alder in western Idaho. Pages 111-123 *in* Proceedings of Terrestrial and Aquatic Vegetation. Eastern Washington State College, Cheney, WA.

Miller TB. 1976. Ecology of riparian communities dominated by white alder in western Idaho. Moscow (ID): University of Idaho, 154 p.

Millspaugh SH, Whitlock C, Barlein PJ. 2000. Variations in fire frequency and climate over the past 17 000 years in central Yellowstone National Park. *Geology* 28 (3), 211 –214.

Montgomery DR, Buffington JM, Smith RD, Schmidt KM, Pess G. 1995. Pool spacing in forest channels. *Water Resources Research* 31:1097-1105.

Montgomery DR, Buffington JM. 1997. Channel reach morphology in mountain drainage basins. Geological Society of America Bulletin 109: 596-611.

Montgomery DR, Buffington JM. 1998. Channel processes, classification, and response. River Ecology and Management, Eds. Naiman R and Bilby R. New York (NY): Springer Verlag.

Montgomery DR, Gran KB. 2001. Downstream variations in the width of bedrock channels. *Water Resources Research* 37:1841-1846.

Morris GL, Fan J. 1998. *Reservoir Sedimentation Handbook*: Design and management of dams, reservoirs, and watersheds for sustainable use. New York (NY): McGraw-Hill.

Moscrip A, Montgomery DR. 1997. Urbanization, flood frequency, and salmon abundance in Puget Lowland streams. *Journal of the American Water Resources Association*. 33 (6): p. 1289-1297.

Moye FJ, Hacket WR, Blakley JD, Snider LG. 1988. Regional geologic setting and volcanic stratigraphy of the Challis vocalic field, central Idaho. In: Link PK, Hackett WR, editors. *Guidebook to the geology of central and southern Idaho*. Moscow (ID): Idaho Geological Survey, Bulletin 27: p. 87–97.

Naiman RJ, Johnson CA, Kelley JC. 1988. Alteration of North American streams by beaver. *BioScience* 38**:** 753-762.

Naiman RJ. 2001. Aquatic and Fisheries Sciences, University of Washington. Seattle, Washington. Personal communication with Sherrill Doran, CH2MHILL, Boise, Idaho, on May 30, 2001.

NCASI. 1999. Silviculture and water quality: a quarter century of Clean Water Act progress. Special Report No. 99-06. Corvallis (OR): National Council for Air and Stream Improvement. December 1999.

Nelson RL, McHenry ML, Platts WS. 1991. Mining. In: Meehan WR, editor. Influences of forest and rangeland management on salmonid fishes and their habitats. Bethesda (Maryland): *American Fisheries Society Special Publication* 19.

NPPC. 2000. *Return to the River*. Portland (OR): Northwest Council, Northwest Power Planning Council, Publication 2000-12.

NPS. 1980. Final Wild and Scenic River Study Report and Environmental Statement, Snake River. Seattle (WA): UDDI, NPS.

NRCS. 1995. Sedimentation in Irrigation Water Bodies, Reservoirs, Canals, and Ditches. Working Paper No. 5, retrieved from http://www.nhq.nrcs.usda.gov/RCA_PAPERS/WP05/wp05text.html on September 10, 2001.

O'Brian P. 2001. UACE, Portland, Oregon. Personal communication with Kelvin Anderson, Idaho Power Company, Boise, Idaho on December 1, 2001.

O'Connor JE, Grant GE, Curran JH, Fassnacht H. 1999. Geomorphology of the Deschutes River below the Pelton Round Butte Dam Complex, Oregon. April 1999.

O'Connor JE. 1993. Hydrology, hydraulics, and geomorphology of the Bonneville Flood. Boulder (CO): Geological Society of America, Special Paper 274.

Osterkamp W. 1997. Expert witness report in SRBA Case No. 39576. Twin Falls (ID): Prepared for the U.S. Fish and Wildlife Service.

Othberg KL, O'Connor JE, McDaniel PA. 1996. Field guide to the Quaternary geology of the Boise Valley and adjacent Snake River Valley. Moscow (ID): Idaho Geological Survey, Staff Report 96-1.

Othberg KL. 1994. Geology and geomorphology of the Boise Valley and adjoining areas, western Snake River Plain, Idaho. Moscow (ID): Idaho Geological Survey, Bulletin 29.

Ott JS. 1997. Clearing the country: a history of the Hudson's Bay Company's fur desert policy. University of Montana, Missoula, MT. 127 p.

Palmer T. 1991. *The Snake River: Window to the West*. Washington (DC): Island Press.

Parker G. 1990. Surface-based bedload transport relation for gravel rivers. *Journal of Hydraulic Research* 28:417-436.

Parkinson SK (Ed.). 2001. Project hydrology and hydraulic models applied to the Hells Canyon Reach of the Snake River. *In*: Technical appendices for the Hells Canyon Complex Hydroelectric Project. Technical Report E.1-4. Boise (ID): Idaho Power Company.

Parkinson SK, Anderson K, Conner J, Milligan J. 2003. Sediment transport, supply and stability in the Hells Canyon Reach of the Snake River. *In*: Technical appendices for the Hells Canyon Complex Hydroelectric Project. Technical Report E.1-1. Boise (ID): Idaho Power Company.

Pearson RH. 1978. A dendrochronological study in south-central Idaho. Pocatello (ID): Idaho Museum of Natural History, Archaeological Reports of the Idaho State University Museum, Bulletin 15.

Pierce KL, Scott WE. 1982. Pleistocene episodes of alluvial-gravel deposition, southeastern Idaho. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 685–702.

Platts WS. 1991. Livestock grazing. Influences of forest and rangeland management on salmonid fishes and their habitat. *Amer Fisheries Soc Spec Publ* 19:389-423.

Powers MC. 1953. A new roundness scale for sedimentary particles. *Journal of Sedimentary Petrology* 23:117-119.

Reid KC. 1991. *Prehistory and Paleoenvironments at Pittsburg Landing: Data Recovery and Test excavations at Six Sites in Hells Canyon National Recreation Area, West Central Idaho.* Project Report Number 15, Center for Northwest Anthropology. Pullman (WA): Washington State University.

Reid KC. 1998. State Archaeologist, Idaho State Historic Preservation Office & Archaeological Survey. Personal communication with Mark Druss, IPC, Boise, Idaho in 1998.

Reid LM, Dunne T. 1996. Rapid evaluation of sediment budgets. Reiskirchen, Germany: Catena Verlag.

Reidel SP. 1982. Stratigraphy of the Grande Ronde Basalt, Columbia River Basalt Group, from the Lower Salmon River and Northern Hells Canyon Area, Idaho, Oregon, and Washington. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 77–101.

Rhodes JJ, McCullough DA, Espinosa FA. 1994. A coarse screening process for evaluation of the effects of land management activities on salmon spawning and rearing habitat in ESA consultations. Technical Report 94-4. Portland (OR): Columbia River Inter-Tribal Fish Commission.

Richter BD, Baumgartner JV, Powell J, Braum DP. 1996. A method for assessing hydrologic alternation within ecosystems. *Conservation Biology*, 10(4): 1163-1174.

Richter BD. 2001. Indicators of Hydrologic Alternation: User's manual. The Nature Conservancy with Smythe Scientific Software. July 2001.

Rosgen DL, Silvey HL. 1996. *Applied River Morphology*. Pagosa Springs (CO): Wildland Hydrology.

Rosgen DL. 1994. A classification of natural rivers. *Catena*, Vol. 22: 169-199.

Schmidt JC, Rubin DM. 1995. Regulated streamflow, fine-grained deposits, and effective discharge in canyons with abundant debris fans. In Costa JE, Miller AJ, Potter KW, Wilcock PR (eds.) Natural and Anthropogenic Influences in Fluvial Geomorphology. American Geophysical Union, Monograph 89, Washington, DC, pp. 177-196.

Schubert FC. 1903. Report of Survey--Snake River-Imnaha to Lewiston, Idaho. Report to W.C. Langfitt. Engineer Office, U.S. Army, War Department. March 7. In file Upper Columbia and Snake River, Record group 77, 1899-1903, box 210, vol. 1. National Archives, Pacific Northwest Region, Seattle.

Schumm SA. 1977. The idealized fluvial system. *The Fluvial System*. New York (NY): John Wiley and Sons.

SCS. 1980b. Soil survey of Malheur Country, Oregon, Northeastern Part. Portland (OR): U.S. Department of Agriculture.

Shields A. 1936. Anwendung der Aehnlichkeitsmechanik und der Turbulenzforschung auf die Geschiebebewegung. *Mitteilungen der Preussischen Versuchsanstalt fur Wasserbau und Schiffbau* 26:26.

Simons and Associates. 2001. Revised draft: Geomorphic Classification, Hells Canyon. Prepared for Idaho Power. Fort Collins (CO). July 2001.

Simons DB, Sentürk F. 1992. *Sediment Transport Technology*. Littleton (CO): Water Resources Publications.

Skinner BJ, Porter SC. 1987. *Physical Geology*. New York (NY): John Wiley and Sons.

Smathers RL, King BA, Patterson PE. 1995. Economics of surface irrigation systems. Cooperative Extensive System, Bulletin No. 779. Moscow (ID): University of Idaho.

Soil Conservation Service [SCS]. 1980a. Soil survey of Ada County area, Idaho. Boise (ID): U.S. Department of Agriculture.

Spence BC, Lomnicky GA, Hughes RM, Novitzki RP. 1996. An ecosystem approach to salmonid conservation. TR-4501-96-6057. Corvallis (OR): ManTech Environmental Research Services Corp.

Summerfield MA. 1991. *Global Geomorphology: An Introduction to the Study of Landforms*. New York (NY): John Wiley and Sons.

Swanson FJ, Berda LE, Duncan SH, Grant GE, Megahan WF, Reid LM, Zimmer RP. 1987. Mass failures and other processes of sediment production in Pacific Northwest forest landscapes. In *Streamside Management: Forestry and Fisheries Interactions*, ed. Salo EO, Curdy TW. Seattle (WA): College of Forest Resources, University of Washington.

Swirydczuk K, Larson GP, Smith GR. 1982. Volcanic ashbeds as stratigraphic markers in the Glenns Ferry and Chalk Hills formations from Adrian, Oregon, to Bruneau, Idaho. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 543–558.

Thomas WO, Lumb AM, Flynn KM, Kirby WH. 1998. Users manual for program PEAKFQ, annual flood frequency analysis using Bulletin 17B guidelines. Washington (DC): U.S. Geological Survey, Water Resources Investigation Report.

Thompson DM. 2001. Random controls on semi-rhythmic spacing of pools and riffles in constriction-dominated rivers. *Earth Surface Processes and Landforms* 26:1195-1212.

Thompson RS, Anderson KH, Bartlein PJ. 1999. Quantitative paleoclimate reconstructions from Late Pleistocene plant macrofossils of the Yucca Mountain region. Denver (CO): US Geological Survey, Open-File Report 99-338.

Thorne CR. 1997. Channel types and morphological classification. In C. R. Thorne, R. D. Hey, M. D. Newson, editors. Applied fluvial geomorphology for river engineering and management, pp. 175-222, Chichester (England): John Wiley and Sons.

Thornton J. Boise Region, U.S. Forest Service. Personal communication with Steve Miller, CH2M HILL, Boise, Idaho in 2001.

Tisdale EW, Hironaka M, Fosberg MA. 1969. The sagebrush region in Idaho: a problem in range resource management. Moscow (ID): University of Idaho. Ag. Exp. Sta. Bulletin 512. 12 p.

Tisdale EW. 1979. A preliminary classification of Snake River canyon grasslands in Idaho. Moscow (ID): University of Idaho. For. Wildl. and Range Exp. Sta. 8 p.

Tisdale EW. 1986a. Canyon grasslands and associated shrublands of west central Idaho and adjacent areas. Moscow (ID): Univ of Idaho. For Wildl and Range Exp Sta Bull 40. 42 p.

Tisdale EW. 1986b. Native vegetation of Idaho. *Rangelands* 8:202-207.

Trimble SW, Crosson P. 2000. "2000 U.S. Soil Erosion Rates - Myth and Reality." *Science*. Vol. 289. July 14, pp. 248-250.

University of Idaho. 1995. Mean annual precipitation for the State of Idaho (1961-1990). Moscow (ID): State Climate Program, compiled in conjunction with U.S. Department of Agriculture, Bureau of Land Management, Bureau of Reclamation, and the Idaho Water Resources Research Institute.

U.S. Army Corps of Engineers [USACE]. 1996. Aerial photographs of Hells Canyon. 1:8400.

USACE. 1999. Lower Snake River Juvenile Salmon Migration Feasibility Study and Environmental Impact Statement, Snake River Flow Augmentation Impact Analysis Appendix. Boise (ID): Prepared for the U.S. Army Corps of Engineers, Walla Walla District by U.S. Bureau of Reclamation, Pacific Northwest Region. February 1999.

U.S. Bureau of Reclamation [USBR]. 1984. Unpublished sediment data for Black Canyon Reservoir collected by USBR in December 1984. Boise (ID): Pacific Northwest Regional Lab.

U.S. Census Office. 1896. Agriculture by irrigation in the western part of the United States at the eleventh census, 1890. House of Representatives Miscellaneous Documents 340 part 20, first session of the fifty-second Congress. Report on the statistics of agriculture in the United States at the eleventh census: 1890. U.S. Government Printing Office, Washington, DC.

U.S. Department of Agriculture [USDA]. 1936. The western range: letter from the Secretary of Agriculture in response to Senate Resolution No. 289. A report on the western range—a great but neglected natural resource. U.S. Senate, 74th Congress, 2nd session. Document 199. Washington (DC): U.S. Government Printing Office.

U.S. Fish and Wildlife Service. 2001. Deer Flat National Wildlife Refuge Home Page. Accessed from http://pacific.fws.gov/deer/ on November 14, 2001. Nampa (ID): U.S. Fish and Wildlife Service, Deer Flat National Wildlife Refuge, November 4, 1997.

USBR. 1994. Management Plan Montour Wildlife/Recreation Area, Montour Valley, Idaho. May 1984.

USBR. 1997. Five Mile Drain: Urbanization of flood hydrology development, Boise, Idaho. (Denver (CO): U.S. Bureau of Reclamation, Technical Service Center. January 1997.

USBR. 1998. Biological Assessment: Bureau of Reclamation Operations and Maintenance in the Snake River Basin Above Lower Granite Reservoir, Boise, Idaho. Boise (ID): U.S. Bureau of Reclamation, Pacific Northwest Region. April 1998.

USBR. 1999. Reservoir Sedimentation Information Worksheet. Denver (CO): U.S. Bureau of Reclamation, Sedimentation and River Hydraulics Group. December 1999.

USBR. 2001. Arrowrock Reservoir sedimentation quantification and transport study report. Denver (CO): Technical Service Center, Water Resources, Sediment and River Hydraulic Group. Revised February 2001.

USDA. 1980. Final environmental impact statement, Hells Canyon National Recreation Area, comprehensive management plan. Washington (DC): Wallowa-Whitman, Nez Perce, and Payette National Forests.

USDA. 1990. Land and Resource Management Plan. Baker City (OR): Wallowa-Whitman National Forest.

USDA. 1991. *Influences of Forest and Rangeland Management on Salmonid Fishes and Their Habitats*. Ed. W.R. Meehan. Bethesda (Maryland): U.S. Department of Agriculture, Forest Service Pacific Northwest Research Station and Forest Environment Research.

USDA. 1995. A comparison of stream materials moved by mining suction dredge operations to the natural sediment yield rates. Letter to the record prepared by Michael Cooley. October 16, 1995.Grants Pass (OR): U.S. Department of Agriculture, Siskiyou National Forest.

USDA. 1996. Status of Interior Columbia Basin - Summary of Scientific Findings. Portland (OR): U.S. Forest Service, Pacific Northwest Research Station, Interior Columbia Basin Ecosystem Management Project. November 1996.

USDA. 1997. South Fork Clearwater Subbasin, Preliminary Report, Landscape Assessment. Grangeville (ID): U.S. Forest Service, Nez Perce National Forest, October 1997.

USDA. 1998. Keys to soil taxonomy, 8th ed. Washington (DC): U.S. Department of Agriculture, Natural Resources Conservation Service.

USDA. 1999a. Soil taxonomy: a basic system of soil classification for making and interpreting soil surveys: 2nd ed. Washington (DC): U.S. Department of Agriculture, Natural Resources Conservation Service, Agricultural Handbook Number 436.

USDA. 1999b. Revised Draft Environmental Impact Statement, Comprehensive Management Plan, Hells Canyon National Recreation Area. Baker City (OR): U.S. Forest Service, Wallowa Whitman National Forest. December 1999.

USDA. 2000. Revised draft environmental impact statement: Hells Canyon National Recreation Area, Comprehensive Management Plan. Baker City (OR): U.S. Department of Agriculture, Wallowa-Whitman National Forest, February 2000.

U.S. Geological Survey [USGS]. 1925. Power resources of the Snake River between Huntington, Oregon, and Lewiston, Idaho. *In*: Contributions to the hydrology of the United States, 1923-1924. Water Supply Paper 520. Washington (DC): U.S. Geological Survey.

USGS. 1940. Discharge and sediment loads in the Boise River drainage basin, Idaho 1939–1940. Washington (DC): Prepared in cooperation with the Flood Control Coordinating Committee, USDA.

USGS. 1982. Guidelines for determining flood flow frequency. Bulletin 17B of the Hydrology Subcommittee, 194 pp. Washington (DC): U.S. Geological Survey.

USGS. 1994. Water Resources Data—Idaho, Water Year 1994. Volume 2. U.S. Geological Survey Water Data Report ID-94-2.

USGS. 1995. Water Resources Data—Idaho, Water Year 1995. Volume 2. U.S. Geological Survey Water Data Report ID-95-2.

USGS. 1997a. Water Resources Data—Idaho, Water Year 1997. Volume 2. U.S. Geological Survey Water Data Report ID-97-2.

USGS. 1997b. Water Resources Data—Oregon, Water Year 1997. Volume 2. U.S. Geological Survey Water Data Report OR-97-1.

USGS. 1999. Water Resources Data, Idaho, Water Year 1999. Washington, D.C.

USGS. 2001a. Magnitude and frequency of instantaneous peak flow at gaging stations in Idaho. Retrieved from http://idaho.usgs.gov/swdata/floodfreq.html on January 11, 2001.

USGS. 2001b. Historical flow and water quality parameter data collected at gauging stations in Idaho. Retrieved from http://water.usgs.gov/id/nwis on June 5, 2001.

U.S. Secretary of Agriculture. 1936. The western range: Letter from the Secretary of Agriculture in response to Senate Resolution No. 289. A report on the western range—a great but neglected natural resource . US Senate, 74th Congress, 2nd session. Document 199. Washington (DC): U.S. Government Printing Office.

Vallier TL. 1994. Geologic hazards in the Hells Canyon region of Idaho, Oregon, and Washington. Washington (DC): U.S. Geological Survey, Open File Report 94-213.

Vallier TL. 1998. Islands and rapids: A geologic story of Hells Canyon. Lewiston (ID): Confluence Press.

Webb RH, Melis TS, Griffiths PG, Elliot JG. 1999. Reworking of aggraded debris fans. In Webb RH, Schmidt JC, Marzolf GR, Valdez RA (eds.) The Controlled Flood in the Grand Canyon. Washington (DC): American Geophysical Union, Monograph 110, , pp. 37-52.

Webster GD, Kuhns MJP, Waggoner GL. 1982. Late Cenozoic gravels in Hells Canyon and the Lewiston Basin, Washington and Idaho. In: Bonnichsen B, Breckenridge RM, editors. *Cenozoic Geology of Idaho*. Moscow (ID): Idaho Bureau of Mines and Geology, Bulletin 26: p. 669–683.

Western Regional Climate Center [WRCC]. 2001. Climate summaries for Idaho. Reno (NV): Western Regional Climate Center, Desert Research Institute. Retrieved from http://www.wrcc.dri.edu/ (April 24, 2001).

Wheeler HE, Cook EF. 1954. Structural and stratigraphic significance of the Snake River capture, Idaho-Oregon. *Journal of Geology*: Vol. 62, no.6, pp. 525-536.

White DL, Vallier TL. 1993. Geologic Evolution of the Pittsburg Landing area, Snake River Canyon, Oregon, Idaho and Washington: Statigraphy and Resources. Washington, D.C. U.S. Geological Survey Professional Paper 1439: p. 55-73.

Whiting PJ, Dietrich WE. 1990. Boundary shear stress and roughness over mobile alluvial beds. *Journal of Hydraulic Engineering* 116:1495-1511.

Williams GP, Wolman MG. 1984. Downstream effects of dams on alluvial rivers. Washington (DC): U.S. Geological Survey, Professional Paper 1286.

Wischmeier WH, Smith DD. 1978. Predicting rainfall erosion losses. Agricultural Handbook 537. Washington (DC): U.S. Department of Agriculture, Agricultural Research Service.

Wohl EE, Thompson DM, Miller AJ. 1999. Canyons with undulating walls. *Geological Society of America Bulletin* 111:949-959.

Woodsmith RD, Buffington JM. 1996. Multivariate geomorphic analysis of forest streams: Implications for assessment of land use impact on channel condition. *Earth Surface Processes and Landforms* 21:377-393.

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Tables

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Table 1.1. Summary of Dominant Lithologies by Drainage Basin (acres)

Basin	Alluvial/ Eolian	Calc-Alkaline Intrusive	Felsic Volcanic	Mafic Volcanic	Meta Sedimentary	Other Volcanic	Other
Upper Snake[a]	7,388,510	526,849	4,764,860	8,207,248	1,338,116	1,199,712	4,695,143
	26%	2%	17%	29%	5%	4%	17%
Owyhee	838,971	35,536	2,079,934	3,172,773	48,351	340,186	540,868
	12%	1%	29%	45%	1%	5%	8%
Boise	520,721	1,602,674	39,285	169,153	3,417	7,623	241,075
	20%	62%	2%	7%	0%	0%	9%
Malheur	260,880	9,545	606,869	1,486,748	13,744	131,719	564,128
	8%	0%	20%	48%	0%	4%	18%
Payette	331,979	1,282,697	92	276,503	19,534	0	217,823
	16%	60%	0%	13%	1%	0%	10%
Weiser	108,059	44,524	0	755,689	119,737	7,155	42,011
	10%	4%	0%	70%	11%	1%	4%
Burnt	45,209	29,600	36,652	132,062	141,212	111,335	209,430
	6%	4%	5%	19%	20%	16%	30%
Powder	182,349	124,307	24,361	346,297	166,440	71,994	179,250
	17%	11%	2%	32%	15%	7%	16%
Hells Canyon[b]	223,886	73,263	478	883,390	140,046	145,337	243,901
	13%	4%	0%	52%	8%	8%	14%
Imnaha	5,879	7,420	0	485,431	18,787	25,970	5,165
	1%	1%	0%	88%	3%	5%	1%
Salmon	741,475	2,688,589	1,786,466	508,652	684,527	543,888	2,014,692
	8%	30%	20%	6%	8%	6%	22%
Grande Ronde	252,142	136,098	41,506	2,050,112	29,569	41,286	73,603
	10%	5%	2%	78%	1%	2%	3%

Notes:
[a]This reach is defined as those basins that are upstream from the Owyhee River confluence (RM 393).
[b]This reach includes local tributaries (with the exception of the Burnt and Powder rivers) between the Weiser River confluence (RM 351) and Lewiston, Idaho (RM 139).

Table 2.1. Summary of Storage Facilities Upstream of Hells Canyon

Mainstem/Tributary Facilities (Tributary Name)	**Owner or Operator**	**Data Source**	**Year Completed**	**Total Storage (acre-feet)**
Jackson Lake Dam and Lake	USBR	A	1916	847,000
Palisades Dam and Reservoir	USBR	A	1957	1,401,000
Henrys Lake (Henry's Fork)	North Fork Reservoir Company	A	1923	90,300
Island Park Dam and Reservoir (Henry's Fork)	USBR	B	1938	135,586
Grassy Lake Dam and Reservoir (Henry's Fork)	USBR	B	1939	15,470
Grays Lake (Willow Creek)	Bureau of Indian Affairs	A	1924	400,000
Ririe Dam and Reservoir (Willow Creek)	Corps of Engineers	B	1977	100,541
Blackfoot Reservoir (Blackfoot River)	Bureau of Indian Affairs	A	1913	413,000
American Falls Dam and Reservoir	USBR	B	1927	1,672,590
Minidoka Dam and Lake Walcott	USBR	A	1906	210,200
Oakley Reservoir (Goose Creek)	Oakley Canal Company	D	1911	77,000
Milner Dam and Lake	Twin Falls Canal Company	A	1905	50,000
Twin Falls	Idaho Power Company	D	1935	955
Shoshone Falls	Idaho Power Company	C	1907	0
Salmon River Canal Reservoir (Salmon Falls Creek)	Salmon River Canal Company	D	1910	260,650
Upper Salmon A	Idaho Power Company	C	1937	0
Upper Salmon B	Idaho Power Company	C	1947	600
Lower Salmon	Idaho Power Company	C	1949	10,900
Mormon (Twin Lakes) (Malad River)	McKinney and Dairy Creeks	A	1908	31,400
Magic Reservoir (Malad River)	Big Wood Canal Company	A	1917	191,500
Fish Creek Reservoir (Malad River)	Carey Valley Reservoir Company	A	1923	12,740
Little Wood A (Malad River)	USBR	A	1939	12,100
Little Wood B (Malad River)	USBR	A	1960	30,000
Bliss	Idaho Power Company	C	1950	8,415
C.J. Strike	Idaho Power Company	D	1952	250,000
Swan Falls	Idaho Power Company	C	1901	7,425
Owyhee Dam and Reservoir (Owyhee River)	USBR	A	1932	1,120,000
Antelope Dam and Reservoir (Owyhee River)	Jordan Valley Irrigation District	A	1935	26,300
Wild Horse Dam and Reservoir (Owyhee River)	Bureau of Indian Affairs	A	1937	173,500
Deer Flat (Lake Lowell) (Boise River)	USBR	A	1908	173,000
Arrowrock Dam and Reservoir (Boise River)	USBR	A	1915	286,600
Anderson Ranch Dam and Reservoir (Boise River)	USBR	A	1950	493,200
Lucky Peak Dam and Reservoir (Boise River)	Corps of Engineers	A	1957	293,100
Warm Springs Dam and Reservoir (Malheur River)	Warmsprings ID	A	1926	192,400
Agency Valley Dam and Beulah Reservoir (Malheur River)	USBR	A	1935	59,000
Willow Creek Dam and (Malheur) Reservoir (Malheur River)	Orchard ID	A	1939	20,000
Bully Creek Dam and Reservoir (Malheur River)	USBR	A	1963	31,600
Black Canyon Dam and Reservoir (Payette River)	USBR	B	1924	44,700

Table 2.1. (Cont.)

Mainstem/Tributary Facilities (Tributary Name)	Owner or Operator	Data Source	Year Completed	Total Storage (acre-feet)
Lake Fork Dam and Reservoir (Payette River)	Lake Fork ID	A	1926	13,165
Deadwood Dam and Reservoir (Payette River)	USBR	A	1931	162,000
Payette Lakes System (Payette River)	The Lake Reservoir Company	A	1944	42,400
Cascade Dam and Reservoir (Payette River)	USBR	A	1948	703,200
Paddock Valley Reservoir (Payette River)	Little Willow ID	A	1949	25,100
Horsethief Reservoir (Payette River)	Idaho Fish and Game	A	1967	4,900
Crane Creek Dam and Reservoir (Weiser River)	Crane Creek Admin Board	A	1920	60,000
Lost Valley Dam and Reservoir (Weiser River)	Lost Valley Reservoir Company	A	1929	10,300
C. Ben Ross Dam and Reservoir (Weiser River)	Little Weiser River ID	A	1936	7,800
Mann Creek Dam and Reservoir (Weiser River)	USBR	A	1967	12,500
Unity Dam and Reservoir (Burnt River)	USBR	A	1938	25,500
Thief Valley Dam and Reservoir (Powder River)	USBR	A	1932	13,300
Mason Dam and Phillips Lake (Powder River)	USBR	A	1968	95,500
Total Upstream Storage from Brownlee Reservoir				10,318,437
Brownlee	Idaho Power Company	C	1958	1,420,000
Oxbow	Idaho Power Company	C	1961	57,500
Hells Canyon	Idaho Power Company	C	1969	170,000
Total Upstream Storage Including HCC				11,965,937

Notes:
Mainstem facilities are list in the far left column in upstream to downstream order. Tributary facilites are indented and italicized. Tributary facilities are sorted by construction date within each major tributary and each major tributary is presented in downstream order.
A. USBR, 1998a. Biological Assessment for Operations and Maintenance in Snake River Basin Above Lower Granite Reservoir, Combined Report. (Boise (ID): U.S. Bureau of Reclamation, Pacific Northwest Region. April 1998.
B. USBR, 1998b. Biological Assessment for Operations and Maintenance in Snake River Basin Above Lower Granite Reservoir, Operations Manual. (Boise (ID): U.S. Bureau of Reclamation, Pacific Northwest Region. April 1998.
C. IPC, 2001d. Operational data, unpublished. Boise (ID): Idaho Power Company.
D. USGS, 2000. Water Resources Data, Idaho, Water Year 2000. Water Data Report ID-00-1 and ID-00-2 -- Volumes 1 and 2.

Table 2.2. Summary of Study Area Land Use (1998)

Land Use Classification	Percent Total Basin Area
Rangeland	59
Forest lands	23
Cropland and pasture	13
Bare ground	2
Water	<1
Wetlands	<1
Urban and transportation	<1
Industrial and commercial	<1
Mines, quarries, gravel pits	<1
Other agricultural land	<1
Transitional	<1
Perennial snowfield	<1

Table 4.1. Slope Classifications for the Snake River Basin (percentage of subbasin area)

Subbasin	>40 Degrees	30–40 Degrees	10–30 Degrees	<10 Degrees
Upper Snake River[a]	9.4	7.9	24.7	58.1
Owyhee River	2.9	4.7	25.8	66.7
Boise River	24.9	15.4	26.2	33.6
Payette River	21.9	15.9	33.8	28.5
Malheur River	3.4	8.6	48.5	39.6
Weiser River	7.2	12.1	46.7	33.9
Burnt and Powder Rivers	11.8	14.0	47.0	27.2
Hells Canyon[b]	22.1	15.0	39.9	23.0
Imnaha River	45.5	15.9	26.5	12.1
Salmon River	44.2	19.1	26.7	10.0
Grande Ronde River	21.7	12.0	37.8	28.5

Notes:
[a]This reach is defined as those basins that are upstream from Owyhee River confluence (RM 392.6).
[b]This reach is defined as including local tributaries (with the exception of the Burnt and Powder rivers) between the Weiser River confluence (RM 351.8) and Lewiston, Idaho (RM 139).

Table 4.2. Summary of Discharge Data for Major Tributaries and Mainstem Snake River (RM 392-351)

Location (USGS gage, period of record)	Mean Annual Discharge (cfs)	Mean Minimum Discharge (cfs)	Mean Maximum Discharge (cfs)	Percent of Inflow Between Owyhee and Weiser Rivers
Snake River near Murphy	10,930	6,397	22,467	—
Owyhee River	427	4	4,366	6%
Boise River	1,711	401	5,067	24%
Malheur River	325	46	2,959	5%
Payette River	3,046	730	11,063	43%
Weiser River	1,108	108	9,700	16%
Snake River near Weiser	17,938	8,700	43,565	—

Notes:
Values shown for tributaries are for the most downstream available USGS gage in each watershed.
The post-regulated period of record includes that period that follows the last major storage facility for each tributary, Owyhee River below Owyhee Dam, OR, USGS Gage 13183000 (POR 1938-2000); Boise River near Parma, ID, USGS Gage 13213000 (POR 1971-1997); Malheur River below Nevada Dam, OR, USGS Gage 13233300 (POR 1993-2000); Payette River near Payette, ID, USGS Gage 13251000 (POR 1967-2000); Weiser River near Weiser, ID, USGS Gage 13266000 (POR 1967-2000); and Snake River at Weiser, ID, USGS Gage 13269000 (POR 1926-2000).
[b]No Malheur River instantaneous TSS load value is available from USGS.

Table 4.3. Relief Ratios for the Five Major Tributaries Upstream from HCC

	Max Change in Elevation (feet)	Stream Length (miles)	Relief Ratio (feet/feet)	Drainage Basin Area (square miles)
Weiser	4780	83.84	0.0108	1,693
Payette	5926	154.42	0.0073	3,326
Boise	5156	145.42	0.0067	4,037
Malheur	3656	140.47	0.0049	4,803
Owyhee	3765	305.59	0.0023	11,300

Note: Basins are listed from highest to lowest relief ratios.

Table 4.4. Summary of Dominant Lithologies for Total Drainage Basin and Portion of Basin Downstream from Storage (acres)

Basin	Alluvial/ Eolian	Calc-Alkaline Intrusive	Felsic Volcanic	Mafic Volcanic	Meta Sedimentary	Sandstone	Other	Sum
Owyhee								
Total	838,971	35,536	2,079,934	3,172,773	48,351	262,206	618,848	7,056,620
	12%	1%	29%	45%	1%	4%	9%	
DOS	16,535	0	6,472	40,623	0	48,524	7,070	119,224
	14%	0%	5%	34%	0%	41%	6%	2%
Boise								
Total	520,721	1,602,674	39,285	169,153	3,417	167,297	81,401	2,583,949
	20%	62%	2%	7%	0%	6%	3%	
DOS	471,252	56,886	1,592	97,057	3,417	166,775	9,559	806,540
	58%	7%	0%	12%	0%	21%	1%	31%
Malheur								
Total	260,880	9,545	606,869	1,486,748	13,744	467,716	228,131	3,073,633
	8%	0%	20%	48%	0%	15%	7%	
DOS	183,201	5,996	325,914	644,592	9,374	366,371	150,275	1,685,723
	11%	0%	19%	38%	1%	22%	9%	55%
Payette								
Total	331,979	1,282,697	92	276,503	19,534	124,942	92,881	2,128,629
	16%	60%	0%	13%	1%	6%	4%	
DOS	141,794	0	0	106,857	9,471	122,999	1,669	382,789
	37%	0%	0%	28%	2%	32%	0%	18%
Weiser								
Total	108,059	44,524	0	755,689	119,737	23,708	25,458	1,077,176
	10%	4%	0%	70%	11%	2%	2%	
DOS	88,503	41,317	0	606,928	109,833	23,695	18,999	889,275
	10%	5%	0%	68%	12%	3%	2%	83%

DOS – Downstream of storage (using the most downstream facility on each tributary)

Table 4.5. Summary of TSS Data for Major Tributaries and Mainstem Snake River (RM 393-351)

Location	Average TSS Concentration (mg/L)	Mean Annual Discharge (cfs)	TSS Load Average (tons/day)	TSS Load Instantaneous (tons/day)	Average Annual Load (tons/year/ square mile)
Owyhee River	82	427	95	55	3.1
Boise River	69	1,711	318	367	29
Malheur River	56	325	49	n/a[b]	3.7
Payette River	34	3,046	279	513	31
Weiser River	101	1,108	302	1,593	65
Snake River nr Weiser	62	17,938	2,999	3,620	16

Notes:
Post-regulated discharge values shown are for the most downstream available USGS gage in each watershed (see Tables 2.2 and 4.2).
All average TSS concentrations and instantaneous load calculations are from the USGS (1997) database, with the exception of the Malheur River TSS value, which is from an unpublished IPC database (collected between 1995-1999).

Table 5.1. IHA Results for the Regulated and Unregulated Operational Scenarios

	Current Scenario	Run of River Scenario
Basic Parameters		
Area of River Basin (mi^2)	72,590	72,590
Mean Annual Flow (cfs)	19,925	19,909
Mean Flow/Area	0.27	0.27
Annual C.V.	0.42	0.38
Flow Predictability (%)	0.64	0.65
Constancy/Predictability	0.83	0.87
% of Floods in 60-day period	0.27	0.32
Flood-free season (days)	67	83
Mean Discharge (cfs)		
January	20,835	18,465
February	23,947	21,250
March	25,829	24,996
April	31,911	31,202
May	23,613	30,010
June	24,841	24,847
July	13,987	13,096
August	12,679	11,645
September	17,692	13,408
October	16,638	15,724
November	11,234	16,417
December	16,179	17,940
Minimum/Maximums (cfs)		
1-day minimum	9,527	8,476
3-day minimum	9,635	8,998
7-day minimum	9,750	9,333
30-day minimum	10,119	10,328
90-day minimum	12,311	11,664
1-day maximum	49,256	48,612
3-day maximum	47,447	46,960
7-day maximum	45,018	44,304
30-day maximum	37,870	37,511
90-day maximum	32,020	32,440
Other Metrics		
Number of zero days	0.00	0.00
Base Flow	0.52	0.49
Julian date of minimum	263	216
Julian date of maximum	133	117
Low Pulse Count	0.4	1.7
Low Pulse Duration	2.8	2.3
High Pulse Count	2.7	2.8
High Pulse Duration	13.2	15.9
Low pulse Threshold	8,021	8,011
High pulse Level	31,829	31,808
Rise Rate	921.4	1135.8
Fall Rate	-943.4	-1070.3
Number of Reversals	110.5	162.8

Note: Values for-year period (1928-1999).

Table 5.2. Summary of Distribution of Nearshore Geomorphic Features for Two-Mile Segments

Geomorphic Feature	Mean (%)[1]	Minimum (%)[1]	Maximum (%)[1]
Hillslope	54/54	4/0	98/100
Bar	20/18	0/0	92/90
Fan	18/18	0/0	62/96
Terrace	8/9	0/0	82/61
Island	NA	0/0	NA
Upper Riverbed	<1/<1	0/0	18/13
Tributary	<1/<1	0/0	5/3

Note:
[1]Idaho side of the river/Washington-Oregon side of the river

Table 5.3. Discharge Inputs to the MIKE 11 Model

Location	Gage No.	$Q_{1.5}$ (m^3/s)[a]	Q_{100} (m^3/s)[a]
Snake R. at Hells Canyon Dam[b]	13290450	1,125[c]	3,588
Imnaha R. at Imnaha, OR	13292000	60	443
Salmon R. at White Bird, ID	13317000	1,460	3,595
Grande Ronde R. at Troy, OR	13333000	349	1,231

[a]Based on Log-Pearson Type III analysis of annual peak flows.
[b]Flood frequency analysis for the Hells Canyon station is limited to the post-construction period of record (1968-2001).
[c]For comparison, the $Q_{1.5}$ at Hells Canyon Dam is approximately 32 percent greater than the plant capacity.

Table 5.4. Channel Reaches in Hells Canyon

			Confinement Type		Sinuosity	
Reach	Location	River Mile	Bedrock	Mixed[a]	Low	Moderate
1	Hells Canyon Dam to Bills Ck.	247.7-233.1	X		X	
2	Bills Ck. to Sheep Ck.	233.1-229.3		X	X	
3	Sheep Ck. to Pine Bar	229.3-227.6	X		X	
4	Pine Bar to nr. Pleasant Valley Ck.	227.6-213.4		X		X
5	nr. Pleasant Valley Creek to nr. Copper Ck. Resort	213.4-205.7	X			X
6	nr. Copper Ck. Resort to nr. Big Sulphur Ck.	205.7-200.1		X		X
7	nr. Big Sulphur Ck. to nr. Dug Bar	200.1-196.9	X			X
8	nr. Dug Bar to White Horse Rapids	196.9-194.4		X		X
9	White Horse Rapids to nr. First Ck.	194.4-186.2	X			X
10	nr. First Ck. to nr. Cougar Bar	186.2-179.5	X		X	
11	nr. Cougar Bar to nr. China Garden Rapids	179.5-175		X		X
12	nr. China Garden Rapids to nr. Limekiln Rapids	175-170	X			X
13	nr. Limekiln Rapids to nr. Buffalo Eddy	170-161		X		X
14	nr. Buffalo Eddy to nr. Tenmile Canyon	161-152	X			X
15	nr. Tenmile Canyon to nr. Asotin	152-145		X		X

[a] Mixed bedrock and alluvial terraces.

Table 5.5. Channel-unit Types Identified in Hells Canyon

Channel Unit Type	Notes
Pools	Morphologically distinct depression with visible head, bottom and tail
Self-formed	Turbulent scour due to the interaction of flow and sediment transport in the absence of external controls
Forced	Scour induced by external flow obstructions (bedrock projection, debris fans) or constrictions
bedrock	
tributary debris fan	
landslide debris fan	
talus fan	
debris-flow/avalanche fan	
Shallows	Relatively planar bed topography and relatively shallow flow
glide/run *riffle* *rapids* (see text) *morphologic* *hydraulic*	Increasing slope, grain size, and relative roughness (ratio of grain size to flow depth) as one moves from glide/run to rapids. glides/runs have flow depths approximating pools.
Bars	Depositional sites complimenting pools
alternate/lateral	Flow divergence and sediment deposition on alternating sides of the channel as part of a classic pool-riffle morphology (e.g., Montgomery and Buffington 1997)
medial	Flow divergence and sediment deposition associated with expansion into a relatively wide channel section
eddy	Flow separation and sediment deposition in the lee of a debris fan
Debris Fans	See text
tributary	
landslide	
talus	
debris-flow/avalanche	
Terraces	Bonneville Flood deposits, landslide backwater deposits, relict bars, relict debris fans

Table 5.6. Summary of Hells Canyon Hillslope Analysis by USGS 6th level HUC

ID	USGS HUC 6	Subwatershed Drainage	Total Area (ha)	Slope <10 degrees (ha)	Slope <10 degrees (% area)	Slope 10 to 30 degrees (ha)	Slope 10 to 30 degrees (% area)	Slope 30 to 40 degrees (ha)	Slope 30 to 40 degrees (% area)	Slope 40 to 60 degrees (ha)	Slope 40 to 60 degrees (% area)	Slope > 60 degrees (ha)	Slope > 60 degrees (% area)
1	10804	Deep Creek ID	5,281	56	1.1%	731	13.8%	831	15.7%	2,869	54.3%	794	15.0%
2	10901	Granite Creek	8,613	100	1.2%	894	10.4%	1,144	13.3%	4,781	55.5%	1,694	19.7%
3	10802	Battle Creek, OR Brush Creek, ID	10,000	106	1.1%	631	6.3%	838	8.4%	4,738	47.4%	3,688	36.9%
4	10803	Saddle Creek, OR	4,656	100	2.1%	338	7.2%	531	11.4%	2,838	60.9%	850	18.3%
5	10702	Sheep Creek, ID	5,813	69	1.2%	675	11.6%	738	12.7%	3,206	55.2%	1,125	19.4%
7	10701	Sheep Creek, ID	4,713	6	0.1%	281	6.0%	538	11.4%	3,369	71.5%	519	11.0%
	Subtotal	Sheep Creek, ID	10,525	75	0.7%	956	9.1%	1,275	12.1%	6,575	62.5%	1,644	15.6%
6	10801	Bernard Creek, ID Three Creek, ID Rush Creek, OR Sluice Creek, OR	12,206	125	1.0%	1,044	8.6%	1,156	9.5%	5,531	45.3%	4,350	35.6%
8	10603	Sand Creek, OR	5,094	56	1.1%	544	10.7%	575	11.3%	3,113	61.1%	806	15.8%
9	10602	Kirkwood Creek, ID	3,775	13	0.3%	638	16.9%	856	22.7%	2,075	55.0%	194	5.1%
10	10604	Temperance Creek, OR	3,875	125	3.2%	463	11.9%	869	22.4%	2,119	54.7%	300	7.7%
11	10601	Salt Creek, OR	4,781	81	1.7%	706	14.8%	975	20.4%	2,850	59.6%	169	3.5%
12	10503	Corral Creek, ID Klopton Creek, ID	7,363	150	2.0%	1,475	20.0%	1,850	25.1%	3,456	46.9%	431	5.9%
13	10502	Kurry Creek, ID	6,150	188	3.0%	1,175	19.1%	1,400	22.8%	3,038	49.4%	350	5.7%
14	10501	Big Canyon Creek, ID Somers Creek, OR	6,800	56	0.8%	738	10.8%	1,375	20.2%	4,450	65.4%	181	2.7%
15	10401		5,781	88	1.5%	638	11.0%	1,331	23.0%	3,506	60.6%	219	3.8%
16	11001	Deep Creek, OR	6,950	419	6.0%	1,600	23.0%	1,406	20.2%	3,381	48.7%	144	2.1%
17	10402	Getta Creek, ID	5,094	181	3.6%	1,063	20.9%	1,150	22.6%	2,613	51.3%	88	1.7%
18	10103	Dry Creek, ID	5,163	113	2.2%	1,094	21.2%	1,475	28.6%	2,444	47.3%	38	0.7%
19	10101	Confluence with Salmon River, ID Confluence with Imnaha River, OR	3,806	25	0.7%	525	13.8%	763	20.0%	2,169	57.0%	325	8.5%
20	10102	Dug Creek, OR	5,250	88	1.7%	575	11.0%	1,106	21.1%	3,275	62.4%	206	3.9%
21	10301	Wolf Creek	4,975	131	2.6%	1,406	28.3%	1,094	22.0%	2,194	44.1%	150	3.0%
25	10302	Wolf Creek	5,900	2,181	37.0%	1,750	29.7%	875	14.8%	1,056	17.9%	38	0.6%
	Subtotal	Wolf Creek	10,875	2,313	21.3%	3,156	29.0%	1,969	18.1%	3,250	29.9%	188	1.7%
22	30502	Cherry Creek, OR	4,988	344	6.9%	994	19.9%	988	19.8%	2,613	52.4%	50	1.0%
23	10201	Divide Creek, ID	7,975	738	9.2%	3,356	42.1%	1,644	20.6%	2,181	27.4%	56	0.7%
24	30503	Cook Creek, OR	6,994	394	5.6%	1,131	16.2%	1,225	17.5%	4,038	57.7%	206	2.9%
26	30501		7,950	94	1.2%	913	11.5%	1,881	23.7%	4,631	58.3%	431	5.4%
27	30402	Shovel Creek, OR Lower Cache Creek, OR Big Cougar Creek, ID	8,825	250	2.8%	1,338	15.2%	1,963	22.2%	4,969	56.3%	306	3.5%
28	30403	Corral Creek, ID	2,063	31	1.5%	313	15.2%	594	28.8%	1,119	54.2%	6	0.3%
29	30401	Birch Creek, OR Chimney Creek, ID	4,713	225	4.8%	706	15.0%	981	20.8%	2,531	53.7%	269	5.7%
Total			**175,544**	**6,531**	**3.7%**	**27,731**	**15.8%**	**32,150**	**18.3%**	**91,150**	**51.9%**	**17,981**	**10.2%**

Table 5.7. Hells Canyon Slope Class per Lithology Unit

	Area (hectares) of each slope class					
Slope Range (degrees):	**<10**	**10 to 30**	**30 to 40**	**40 to 60**	**>60**	**Total**
Geology Class						
Columbia River Basalt	4,169	13,881	13,650	32,575	2,188	66,463
Metavolcanic & Metasedimentary - Seven Devils	563	6,656	8,806	32,050	13,338	61,413
Quartz Diorite/Granodiorite/Quartz Monzonite	163	1,194	1,819	5,888	938	10,000
Terrace Gravels	200	344	88	113	6	750
Alluvium	44	63	6	6	-	119
Landslide Deposits	56	200	131	431	219	1,038
Glacial Deposits	—	—	19	188	25	231
Total Hectares of Each Slope Class	5,194	22,338	24,519	71,250	16,713	140,013

	Percent of geology unit of each slope class					
Slope Range (degrees):	**<10**	**10 to 30**	**30 to 40**	**40 to 60**	**>60**	**Total**
Geology Class						
Columbia River Basalt	6.27%	20.89%	20.54%	49.01%	3.29%	100%
Metavolcanic & Metasedimentary - Seven Devils	0.92%	10.84%	14.34%	52.19%	21.72%	100%
Quartz Diorite/Granodiorite/Quartz Monzonite	1.63%	11.94%	18.19%	58.88%	9.38%	100%
Terrace Gravels	26.67%	45.83%	11.67%	15.00%	0.83%	100%
Alluvium	36.84%	52.63%	5.26%	5.26%	0.00%	100%
Landslide Deposits	5.42%	19.28%	12.65%	41.57%	21.08%	100%
Glacial Deposits	0.00%	0.00%	8.11%	81.08%	10.81%	100%
Total Percentage of Each Slope Class	3.71%	15.95%	17.51%	50.89%	11.94%	—

Notes:
Data from GIS coverage entitled “Hellsgeo” provided by IPC on May 17, 2001, and geological groupings determined by CH2M HILL. The “Hellsgeo” coverage not as complete spatially (140K hectares; Figure 1.5) as the ICBEMP database (175K hectares in Table 5.3; Figure 5.3).

Table 5.8. Coarse-grained Snake River Bed Material Lithologies

Rock Groups	Comments	Percentage
Group 1 Basalt	Microscopically pitted rocks have dark green mafic minerals (probably dominated by olivine) on fresh rock surfaces that weather to yellowish-brown to orange (iron-rich) powders. These characteristics are typical of basalts, particularly the un-metamorphosed Columbia River Basalts that are present within the canyon both upstream and downstream from the complex.	41
Group 2 Basalt/Plagioclase	May include debris from the Columbia River Basalt (Imnaha Basalt) but the weathering habit suggests that this group may be dominantly from the metamorphosed basalt in the Wild Sheep Formation.	18
Group 3 Quartzite/Siltstone	Most rocks are from either localized metamorphic or meta-sediment (highly-altered sedimentary) rocks of the Seven Devils Group (Hunsaker Creek and Doyle Creek Formations) These rocks may be adjacent to local intrusives because both the quartzite and siltstone show essentially complete filling of the pores with silica and partial assimilation of the original grains.	20
Group 4 Diorite	Best described as diorite intrusives, which are the common intrusive rock in Hells Canyon. This intrusive rock type is mineralogically distinctive in that it does not contain typically pink-colored potash feldspar (K-feldspar) that is dominant in the more acidic intrusive rocks from the Idaho Batholith.	9
Group 5 Altered Metamorphics	Dominantly aphanitic pale green, and commonly contain chlorite clay and/or epidote. They may represent metamorphosed clay and shale (argillite) within the Seven Devils Group that have been further altered during the intrusion of the diorite.	6
Group 6 Breccia	Varicolored brown to blue and gray rocks that are highly fractured/broken and contain fragmented clasts cemented together with similarly colored silica. Breccias are common in the Seven Devils Group (Windy Ridge, Hunsaker Formations, and Wild Sheep Formation).	4
Group 7 Argillite	Typically maroon red and are part of the metamorphosed clays/shales of the Seven Devils Group meta-sediments.	2

Table 5.9. Adjusted Sand Bar Surface Areas in Hells Canyon from 1946–1968

Date	RM 229.8	RM 227.5	RM 222.4	RM 216.4	RM 208.3	RM 204.3	RM 201.1	RM 196.8	RM 193.8	RM 192.4
1946	4,746	5,822	3,763	4,764	1,806	1,462		966	2,240	2,020
1948			3,825	4,860	1,877					
1949	4,826	5,961		4,836						
1955						1,520	1,519	1,023	2,294	2,063
1961					1,996	1,641	1,626	1,106	2,371	
1964	4,808	5,928	3,795	4,798	1,855	1,500	1,493	967	2,273	2,047
1968						1,518	1,507	1,013	2,287	2,055

NOTE: Units expressed in square meters. The error associated with these surface area values was roughly determined to be within approximately 100 m^2. This error is a result of the variability involved in rectifying and digitizing the boundaries of the sandbars, shadows in the aerial photos, poor aerial photo images, non-delineated sandbars, resolution, texture and the scale of the photos. Therefore, differences of less than 100 m^2 are not significant (Parkinson et al. 2002).

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