SESSION II

EXOGENIC AND ENDOGENIC PROCESSES

AN OVERVIEW OF CLIMATE CHANGE

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We describe briefly here the main mechanisms and time scales involved in natural and anthropogenic climate variability, based on quantitative paleoclimatic reconstructions from natural archives and climate model simulations: the large glacial-interglacial cycles of the last million years (the Quaternary), lasting typically a hundred thousand years, triggered by changes in the solar radiation received by the Earth due to its position around the Sun; the century-long climatic changes occurring during last glacial period and triggered by recurrent iceberg discharges of the large northern hemisphere ice caps, massive freshwater flux to the north Atlantic, and changes in the ocean heat transport. We show the strong coupling between past climatic changes and global biogeochemical cycles, namely here atmospheric greenhouse gases. We also discuss the decadal climatic fluctuations during the last thousand years, showing an unprecedented warming attributed to the anthropogenic greenhouse gas emissions. We show the range of atmospheric greenhouse concentrations forecasted for the end of the 21st century and the climate model predictions for global temperature changes during the 21st century. We also discuss the possible climatic changes at longer time scales involving the possibility of north Atlantic heat transport collapse (possibility of abrupt climate change), and the duration of the current interglacial period.

1. Natural climate variability: glacial-interglacial cycles of the Quaternary

About 21 000 years ago, the sea-level was ~120 m lower than for present-day, due to an intensive build-up of ice on the northern hemisphere continents with huge North American and North European ice sheets [1]. During this period called the Last Glacial Maximum, Greenland temperatures were 20°C colder [2], Antarctic temperatures 9°C colder [3, 4] and European temperatures about 15°C colder than modern annual means [5]. Huge reorganisations of the oceanic circulation (with larger sea-ice cover) [1] and biosphere types (such as steppe replacing forest over large areas of the mid latitudes) [6] resulted in reduced atmospheric greenhouse gas concentrations (200 ppm of CO_2 versus 280 ppm for interglacial periods; 400 ppb of CH_4 versus 700 ppbv for interglacial periods) [7]. Drier and windier conditions induced huge levels of dust in the atmosphere [7, 8]. Climate models estimate that this dramatic cold climate corresponds to a global temperature cooling of 5°C [9].

Typically warm interglacial conditions occur during ~10 ka and their end corresponds to a progressive glacial inception lasting about ~90 ka and culminating to "glacial maximum" conditions. The forcing (factor external to the climate system) triggering changes between glacial and interglacial "modes" of the climate system is the solar radiation received by the Earth, resulting from changes in its orbit around the Sun [10, 11]. Eccentricity, obliquity and equinox precession result in insolation periodicities of 100, 41, and 19-23 ka. The current understanding of glacial cycles lies in sequences of events involving the insolation forcing (Milankovitch theory) and climate system feedbacks involving icy surfaces albedo and greenhouse gases. At the end of an interglacial period, a marked decrease of

summer solar radiation at 65°N induces a summer cooling, amplified by the persistence of snow cover all year round. The resulting change of vegetation (tundra replacing boreal forest) and ocean circulation (more sea-ice) enhances this albedo feedback [12]. The decrease of wetlands both at polar and tropical locations induces a decrease of CH_4 atmospheric concentrations [13]. Both the cooling and the slowdown of the oceanic circulation reduce the atmospheric CO_2 levels [14]. The reduced greenhouse gases (less CO_2 and CH_4, less water vapour in a cold atmosphere) amplify the cooling leading to the buildup of ice sheets and the glacial climatic conditions. By contrast, the deglaciation processes start in the southern hemisphere that warms up, followed after a few centuries by increased atmospheric CO_2 levels [15], destabilisation of the northern hemisphere ice caps and positive greenhouse and albedo feedbacks.

2. Natural climate variability: abrupt climatic changes

During glacial periods, high resolution paleoclimatic records measured on Greenland ice cores have revealed the existence of a succession of 25 abrupt climatic changes (called Dansgaard-Oeschger events) [16]. They correspond to Greenland temperature changes of typically 10°C to 15°C [17] occurring during several decades to centuries and some of them have recently been identified in high resolution speleothem records of European and Chinese climate [18, 19]. In parallel, North Atlantic deep sea records show changes in deep ocean circulation and reveal among the deep sea sediments the presence of ice rafted debris (rocks and mineral fraction) [20, 21]. The simultaneous changes in atmospheric CH_4 concentration [22, 23] reveals the hemispheric scale of the associated climatic changes. The current understanding of these rapid events is an instability of major ice caps in the northern hemisphere, massive iceberg discharges to the north Atlantic, bringing light freshwater and changing the density of north Atlantic surface waters. Reduced deep water convection and reduced heat transport by the ocean induce large-scale cooling in the northern hemisphere and changes in atmospheric circulation including the tropical areas, and at the same time a warming of the southern hemisphere, in response to an oceanic “see saw” [24]. After a while, the ocean circulation starts again, warming the north Atlantic and cooling the southern hemisphere [25, 26]. These rapid changes directly influence European climate as testified by simultaneous changes in pollen distributions [27]. The largest coolings could have been associated with persistent permafrost in south western France as indicated by hiatus in well-dated speleothem growth [18]. The last of these events took place about 8.2 ka BP (during our interglacial period) and corresponds to a cooling of about 2-5°C and lasting about 40 years [28, 29].

3. Natural and anthropic climate variability: last millennium

The instrumental records of meteorological observations enable hemispheric temperature change estimates for about 150 years. Due to a huge level of interannual variability in the climatic parameters, such records remain short to detect slow centennial scale fluctuations or trends. Reconstructions of northern and southern hemisphere temperatures have been compiled for the last thousand years on an annual basis using multiple proxies of climate change calibrated against meteorological records, including specifically long tree ring chronologies [30]. Historical records of environmental conditions are also heavily used for climate reconstruction purposes [31]. The Little Ice Age, from about A.D. 1350 to A.D. 1860, is the last of the centennial scale climatic fluctuations which is well known from historical documentation. Such a slightly colder time period (about 0.6°C colder in the northern hemisphere) corresponds to dramatic environmental changes in vulnerable areas such as Alpine high altitude places, where glaciers advanced significantly [32]. These small temperature changes can be associated with large changes in the hydrological cycle. Several studies have revealed a dramatic change in the seasonal precipitation cycle at the end of the Little Ice Age in France. A study conducted

on tree ring cellulose isotopes in western France has shown that during the past four centuries, the frequency of summer droughts doubles when mean summer temperatures increase by half a degree [33]. At these time scales, the external forcings of climate are mainly solar irradiance, stratospheric volcanic eruptions, land-use and anthropogenic greenhouse gas and sulfate aerosol emissions [34]. Only since about 1985 does the radiative forcing of greenhouse gases overwhelm the impacts of natural forcings.

Human activities modify the land surface (for instance through deforestation and agriculture, building of large cities, …), the land-amosphere moisture and energy exchanges, which may have large scale impacts on climate [34]. The intensive use of fossil fuels results in dramatic increases of atmospheric greenhouse gas concentrations (with long residence times: ~10 years for CH_4, ~100 years for CO_2), and in sulfate aerosols. Sulfate aerosols have short residence times in the atmosphere, a direct radiative effect (back diffusion of solar radiation) and an indirect radiative effect (they act as cloud condensation nuclei). Reconstructions for the last millennium and climate simulations show that the ongoing global warming cannot be explained without taking into account both natural and anthropogenic forcings [34].

4. Future climate changes: next century

About half of the anthropogenic CO_2 emissions accumulate each year in the atmosphere, leading to the observed exponential increase; the other half of the human emissions is stored in the biosphere and the ocean, with decadal to centennial time constants. Even though all greenhouse gas emissions are stopped tomorrow, the atmospheric CO_2 level will continue to rise during several centuries to concentrations of typically 500 ppm [34]. However economists have calculated scenarios of greenhouse gas emissions taking into account fossil fuel availabilities and under different international frames ranging from "virtuous" scenarii where alternative technologies and strong emission reduction policies induce a stabilisation of the emissions; to "business as usual" scenarii where fossil fuels (fuel, gas, coal) are used at the rate of modern economical growth and development. The atmospheric concentration of CO_2 resulting from these calculations lies between 500 and 800 ppm at the end of the 21st century. All climate models therefore forecast a rise of the Earth temperature ranging between 2 to 6°C at this horizon [34]. This change in temperature will be about at the global mean in western Europe (typically 2-3°C in France) and will be associated with changes in precipitation seasonality in Europe (more rainfall in winter and spring and significantly less in summer). This could result in an increased frequency of winter floods and summer droughts during the 21st century. Sea-level should also increase by several tens of centimeters due to thermal dilatation and melt of small glaciers [34].

5. Future climate changes: next thousands of years

During the next millennium, abrupt changes in north Atlantic thermohaline circulation may take place in response to increased freshwater supply through intense cyclogenesis and rainfall on this key area; or due to melting of the Greenland ice cap [35], which would result in a 6 m increase in sea level. Within a fast global warming context, local abrupt events such as the collapse of the thermohaline circulation may induce a local cooling affecting the North Atlantic area and therefore Europe.

At longer time scales, the Earth orbital parameters will again play a key role in the pacing of glacial inception. The summer insolation at 65°N does not exhibit clear minima during the next 50 000 years. If the mechanisms operating during the Quaternary remain the same for the next thousands of years, it is not expected that any glacial period starts within the next 50 thousand years [36]. Taking into account anthropogenic emissions of CO_2 and its fate in the climate system

(ocean pumping active at 10^2 years, carbonate processes active at 10^3 years and geological silicate erosion processes active at 10^4 years), simplified climate models have shown that the human induced greenhouse gases levels should prevent the onset of the next glaciation until 200 000 years in the future, resulting in the longest warm interglacial periods ever known during the last million years.

Figure 1. The climate system

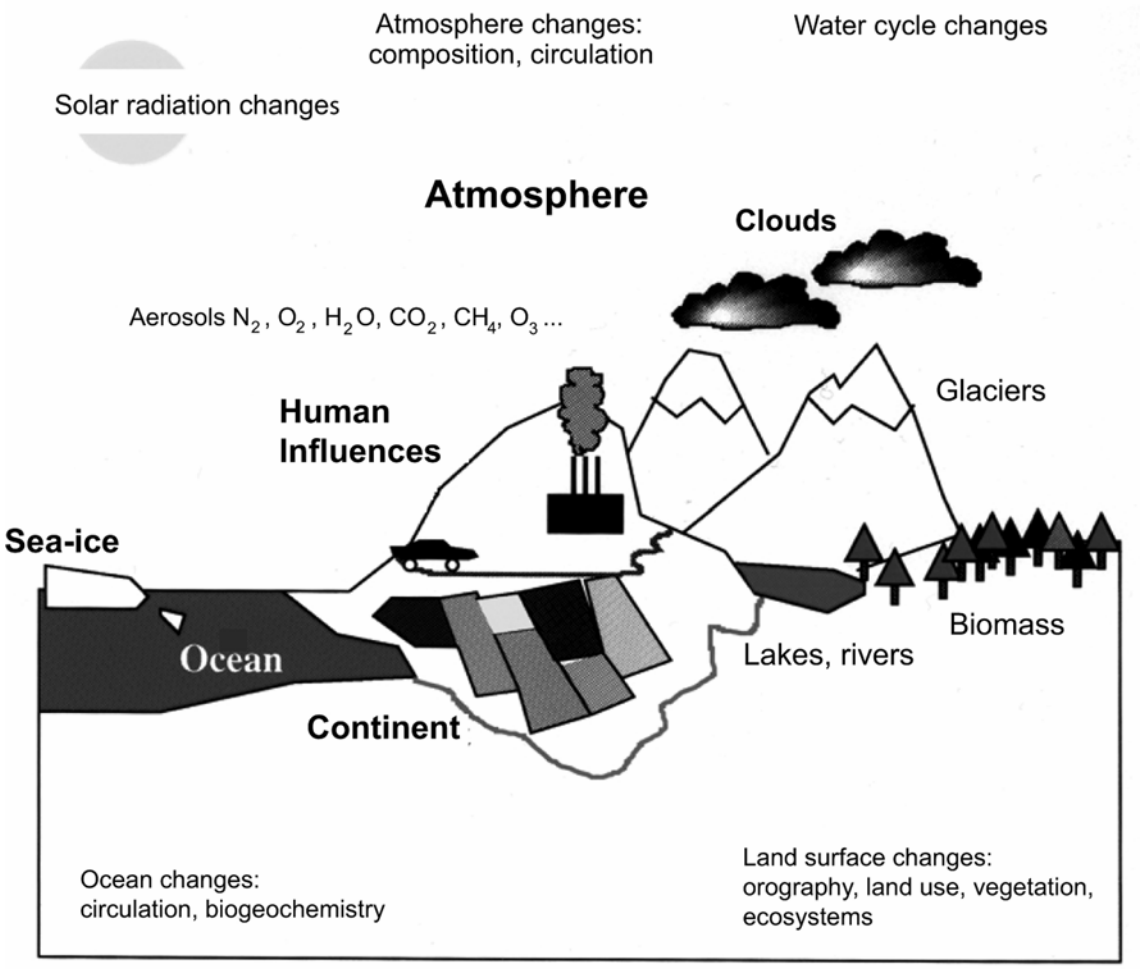
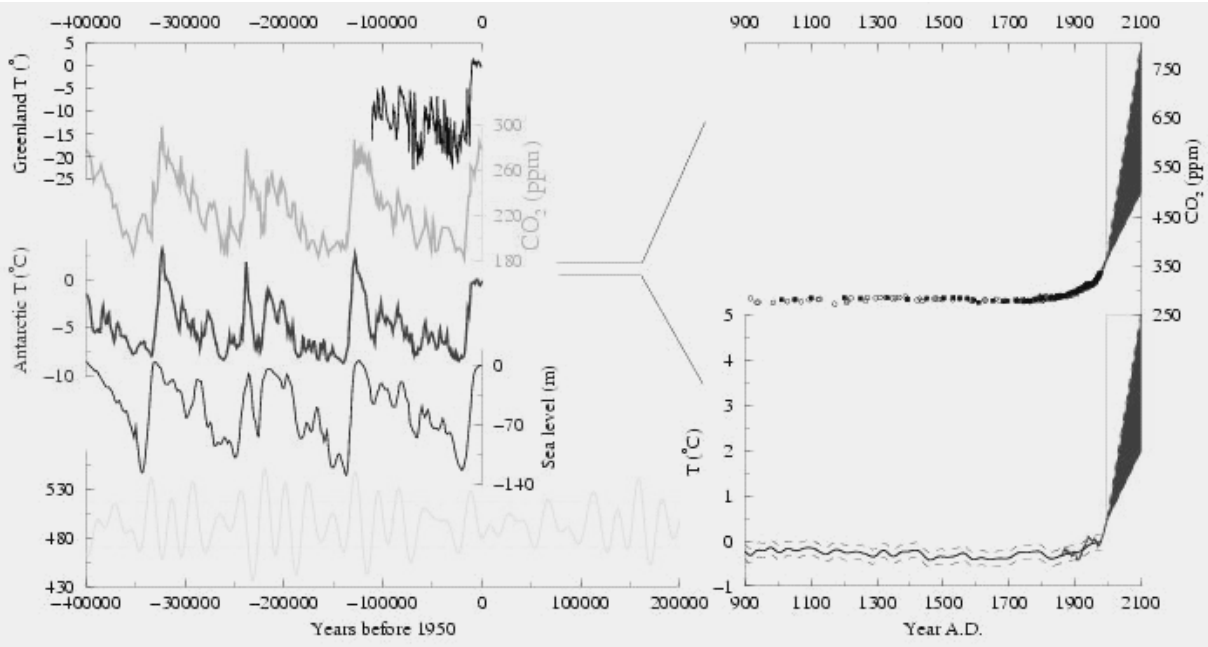


Figure 2. Four glacial-interglacial cycles



(top left panel)	Greenland temperature derived from GRIP ice core $\delta^{18}O$ corrected for changes in snowfall seasonality [37].
(second left panel)	Atmospheric CO_2 fluctuations from Vostok ice cores [7].
(third left panel)	Antarctic temperature derived from Vostok δD [4].
(fourth left panel)	Ice volume recorded by the $\delta^{18}O$ of benthic foraminifera from deep sea sediments [38].
(fifth left panel)	Past and future changes in incoming summer (June 21st) solar radiation at 65°N [39].
(top right panel)	CO_2 fluctuations during the last millennium [40] and projections for the ongoing century [34].
(bottom right panel)	Climatic changes during the last millennium [30] and projections for the ongoing century [34].

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SURFACE EROSION IN OROGENIC SYSTEMS

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1. Introduction

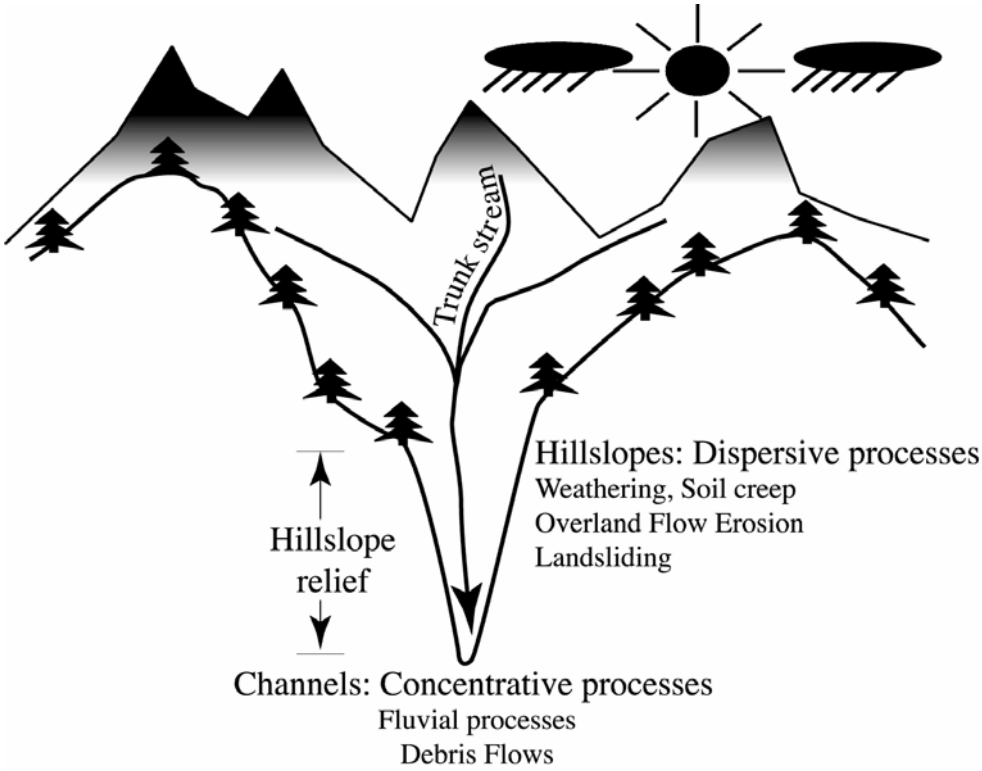
Understanding the effects of climate and tectonics on the geomorphic evolution of mountain belts has become the major research objective in surface processes. In addition, studies have been directed towards exploring possible controls on the local and the global surface mass flux. These objectives can only be addressed if the different components of sediment routing systems are known, and if possible feedbacks between the various erosional mechanisms have been analysed. Accordingly, in a first step, this paper illustrates the nature of surface erosion for different spatial and temporal length scales. In addition, it explores how the landscape evolves from initially simple conditions to more mature stages, and how and to what extent the different components of surface processes influence the topographic development. If this issue has been discussed from a more general point of view, the paper will illustrate how and to what extent climate and crustal processes influence different erosional components and topographic length scales. Finally, it will present quantitative data on current rates of surface erosion for the situation of the Swiss Alps.

2. The nature of landscape evolution

It is generally accepted that except from glacial erosion, long-term (>1 000 years) erosional processes consist of a concentrative and a dispersive component (Figure 1). The concentrative nature of erosion and sediment transport comprises fluvial processes and debris flows in channels. Erosion and sediment transport by fluvial processes then occurs if critical flow strengths are exceeded. In contrast, sediment is transported as debris flows if slopes and pore pressures surpass specific thresholds. The dispersive component of erosion and sediment transport comprises processes on hillslopes such as soil creep that is often referred to as hillslope diffusion in modeling studies (e.g. Tucker & Slingerland, 1997), overland flow erosion and landsliding if critical values for slopes and/or pore pressures are exceeded (e.g. Anderson, 1994).

Despite these general agreements, there are differences in interpretations of how and to what extent these components drive the evolution of landscapes. Some researchers consider fluvial erosion to exert the first-order control because fluvial processes are anticipated to set the lower boundary conditions for geomorphic processes on the adjacent slopes. Accordingly, increasing (or decreasing) rates of fluvial erosion result in equivalent modifications of the local relief and hence in enhanced (or reduced) rates of erosion on the adjacent hillslopes. Alternatively, it was proposed that rates of hillslope erosion do not necessarily have to be controlled by erosion in channels. For instance, Hovius *et al.* (1998) found that hillslope mass wasting at the channel heads govern the rates and the mode of drainage basin modification in the Finisterre Mountains, Papua New Guinea.

Figure 1. Overview of erosional processes



Numerical experiments and field studies carried out in humid (Central Alps, Switzerland) and arid/hyperarid climatic settings (western escarpment of the Andes of northern Chile and Peru) reveal that the controls on surface processes significantly depend on the ratio D_e between the concentrative (fluvial and debris flows) and dispersive (hillslope) characteristic time scales, i.e. on the relative strength of concentrative and dispersive processes (Simpson & Schlunegger, 2003). These studies suggest that fluvial and debris flow dominated systems (high D_e) are characterised by rough, high-relief, highly incised surfaces which contain a dense and hence a closely-spaced channel network, a thin regolith cover and highly sinuous valleys. A classical example is the Napf area of Central Switzerland. Geomorphologies resulting from low D_e-values generally have a thick regolith cover. They display smooth topographies, low reliefs, greater channel spacing and less sinuous valleys. Examples are the densely vegetated regions of the western escarpment of the northern Peruvian Andes.

3. Topographic evolution

The topographic evolution of landscapes tends to occur in two major phases reflecting formation of new channels and valleys, and formation of relief, and a later decay phase related to the gradual destruction of channels and valleys. During this initial growth phase the location of fluvial incision rapidly propagates headward thereby increasing the local relief and the topographic roughness. Besides the modification of relief the phase of backward erosion is characterised by the formation of well-recognisable knick-points in the longitudinal stream profiles, separating landscapes that still record the initial (usually smooth) topography in the headwaters from a geomorphology where the

surface area and the topographic roughness grows as the erosional front passes. Examples can be found in the Andes of Chile and Peru, and in the Swiss Alps. Specifically, in the Andes, enhanced rates of rock uplift at ca. 8 Ma initiated a phase of backward erosion and relief rejuvenation. Similarly, in the Swiss Alps, the base level drop of >80 m that was caused by the retreat of the Alpine glaciers at ca. 15 Ka initiated a phase of downcutting and backward erosion of the Alpine rivers, causing the local relief to grow (Schlunegger & Hinderer, 2003). At these knick-points effective erosion rates are up to 50 times higher than the overall denudation rates in the Alpine and Andean drainage basins.

Theoretical models imply that the relative strength of concentrative and dispersive processes, i.e. D_e, also exerts a significant control on the topographic evolution of mountain belts (Simpson & Schlunegger, 2003). An increase in D_e (which is equivalent to an increase in rates of fluvial erosion and sediment transport) results in a decrease in the timescale within which the topography develops. In this case, the topographic development is characterised by increasing rates of backward erosion and vertical incision, and by an augmentation of the topographic roughness and the local relief. In contrast, an increase in the importance of hillslope processes on the geomorphic development (decrease in D_e) results in a smoothing of the topography and in a decrease in the local relief, the overall denudation rates and the associated surface mass flux.

4. The influence of climate on surface erosion

Climate influences the geomorphic development by production of unconsolidated material through weathering and landsliding, and by transport of sediment through linear processes (such as debris flows and fluvial processes). Observations in the Andes imply that a warm and humid climate enhances weathering rates, which, in turn, results in an increase in the relative importance of dispersive surface processes and hence in a decrease in the topographic roughness, the magnitudes of the local relief and the density of channels and valleys. In contrast, it has been reported that the nature and rates of surface erosion changes significantly if climate becomes more arid. Specifically, a shift towards an arid climate tends to increase the ratio between large- and small-magnitude floods (Turcotte & Green, 1993). In a conceptual publication about climate change and surface eroison, Molnar (2001) thought that an increase in this ratio results in accelerated erosion. The rational behind this argumentation lies in the assumption that a river needs to transport its bedload before the bedrock becomes eroded, and that critical flow strengths must be exceeded to remove this bedload. It appears then that an increase in the frequency of large floods (caused by a shift towards a more arid climate) tends to increase the magnitudes and the frequencies at which these critical thresholds are surpassed. Molnar (2001) then concluded that such a scenario tends to increase the surface mass flux.

Field observations suggest that an increase in aridity tends to decrease weathering rates and hence production rates of regolith. According to the theoretical concepts outlined above, such a climatic shift is anticipated to result in an increase in the relative importance of concentrative (debris flows, fluvial erosion and sediment transport) versus dispersive processes (hillslope processes) and hence in an increase in the channel density, the topographic roughness and the local relief. Such a shift in the erosional regime will also decrease the topographic length scale of the system and hence the rate at which the topography responds to changes in external perturbations. The conclusion of Molnar that an enhanced aridity will increase the erosional mass flux is thus consistent with the results of numerical models even if they are based on different conceptual approaches. In contrast, however, preliminary geochemical and sedimentological data from the western escarpment of the Andes of Chile imply that denudation rates increase with precipitation rates. This increase, however, is currently being documented at a hillslope scale. This does not mean that rates of sediment also increase at the scale of whole drainage basins. Further research is needed to explore this open question.

5. The influence of rock uplift on surface erosion

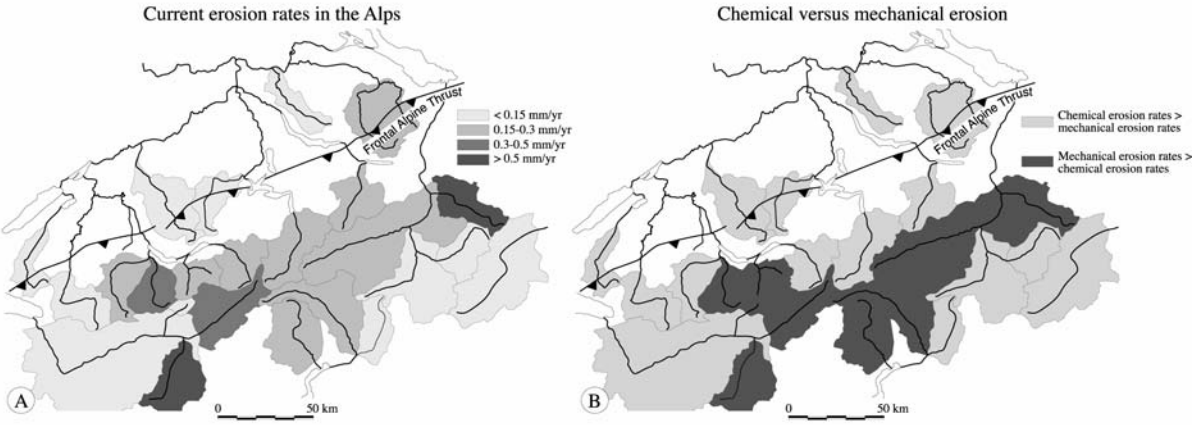
Rock uplift influences erosional processes by an increase in the gradients. Observations from the Andes and from the Central Alps reveal that the erosional systems respond to an increase in rock uplift rates (which is equivalent to a lowering of the base level, or by an overall steepening of the topography) by headward erosion. This phase of backward erosion is characterised by the presence of well-identifiable knick-points in the stream profiles. Beneath the knick-points, the drainage basins have adjusted to the modified pattern of rock uplift. The areas surrounding the knick-points currently experience a phase of relief-rejuvenation as incision has resulted in growth of the local relief. The geomorphologies in the headwaters, in contrast, still record the situation prior to the enhancement of rock uplift. It appears, therefore, that any increase in rock uplift rates (or in base level lowering) will ultimately increase the local relief especially in the segment of active backward erosion. As backward erosion proceeds the location of relief-rejuvenation shifts towards the headwaters. In contrast a reduction in the rates of rock uplift will result in a decrease in fluvial process rates, which, in turn, reduces the ratio between the concentrative and dispersive processes. The result is an overall decrease in the topographic roughness and the local relief.

6. Modern erosion rates in the Central Alps and the importance of bedrock lithology

Modern denudation rates are determined by measurements of sediment loads (mechanical and chemical) of rivers in relation to the size of the whole drainage basins. The data show that total denudation rates increase from ca. 0.15 mm/yr in the distal foreland to ca. 0.5 mm/yr in the Central Alps. Exceptionally high denudation rates of >0.5 mm/yr are observed for drainage basins that are heavily glaciated, and for systems that contain source rocks with low erosional resistance (e.g. Flysch units) (Figure 2A). The tendency towards enhanced average surface erosion from the foreland to the Alpine core is mainly due to the enhanced contribution of mechanical denudation. In contrast, total denudation rates calculated for the drainage basins in the foreland are dominated by chemical denudation reaching up to 90% of the total. Similarly, erosion rates by chemical denudation increase with the relative abundance of carbonates in the source area. The sediment loads imply mean chemical denudation rates of ca. 0.085 mm/yr for systems with predominantly crystalline sources, and rates of ca. 0.125 mm/yr, and 0.105 mm/yr, with catchments made up of carbonates and mixed lithologies, respectively (Figure 2B).

The sediment yields presented in Schlunegger & Hinderer (2003) (Figure 2A) have been calculated based on the prerequisite that erosion operates on the whole surface. However, detailed mapping in Central Switzerland revealed that post-glacial and modern erosion have only modified 30-50% of the surface that formed during glacial times (i.e. 50-70% of the glacial landscape has not been affected by erosion). Specifically, Schlunegger & Hinderer (2003) revealed that modern erosion appears to operate predominantly in V-shaped scars and on the associated hillslopes. In addition, the analysis of digital elevation models in combination with mapping suggest that the ca. 50-70% estimate of non-eroded glacial surface also applies to other portions of the Swiss Alps. Consequently, considerations that chemical erosion affects the whole drainage basin, and mechanical erosion occurs by rivers in the V-shaped scars will result in effective modern incision rates that are for some basins >50% higher than the average yields of Figure 2A.

Figure 2. (A) Total modern erosion rates for selected drainage basins, and (B) the relative importance of chemical dissolution on the erosional mass flux (Modified, Schlunegger & Hinderer, 2003)



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FAULTING AND HAZARD IN LOW SEISMICITY AREAS

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Extended abstract

The relationship between faulting and seismicity in areas of low to moderate seismicity, such as Northern Europe, tends to be obscure. Thus, concepts in hazard analysis originating in the context of active tectonic environments such as California may not be readily applicable. A classic example is the search for "active" faults, where the standard definition of an active fault follows from that laid down by USEPA (1981) along the lines that any fault that can be shown to have produced an earthquake in the last 10 000 years is so categorised. Although one can, and in some circumstances must, evaluate a list of local mapped faults to assess their seismic capability, there is a danger of overlooking the fact that all earthquakes must occur on some fault. Thus, in a country like the UK where around 300 events are detected every year, the number of faults that are "active" by the USEPA definition must also be in the hundreds, although one is in the position of not being able to name a single one of them with complete confidence.

As a result, focussing on a few cases where one can, with some reasonably high probability, suppose individual faults to be associated with certain earthquakes, doing so runs the risk of presenting a skewed picture of the seismicity, by treating some earthquakes differently from others because they can be connected to a mapped fault, whereas in seismological terms, an earthquake that cannot be so connected nevertheless has the same physical properties.

There is some merit in abandoning the idea of active and inactive faults in favour of a three-way classification, more suited to environments in which most fault movement occurs as reactivation of extremely old structures in a tectonic environment different from that in which they originally formed.

Inert faults are those that appear to have no possibility of reactivation in the present stress regime. Such faults have been classed as "extinct" by Muir Wood and Mallard (1992), but the term "inert" avoids inapt comparisons with volcanology.

Reactivatable faults are those that could be reactivated or have been reactivated. Reactivation is seen as a diffuse and largely disorganised process within a particular crustal volume.

Controlling faults are those that are the main expressions of continuing deformational activity, e.g. graben-bounding faults in areas of present extensional tectonics.

Both of the last two categories may be "active" in the sense of hosting modern earthquakes, but a fault in the latter category can be seen as the preferential locus for future movement, and thus future seismic activity along it is certain to occur. Thus, in the UK, the Lake District Boundary Fault, was, during the Permian and Early Triassic, a controlling fault for the subsidence of the East Irish Sea Basin (Akhurst *et al.,* 1998). Today, it may be subject to reactivation, given the occurrence of some small

earthquakes within its traces (Musson, 1987; 1998), but it does not seem to occupy a role more significant than any other fault in the North of England, where many earthquake epicentres may be found.

The assessment of faulting and seismicity has importance in two respects to hazard for nuclear waste disposal, depending on whether one is concerned with rupture (displacement) hazard or vibratory hazard. Both issues need to be addressed; they also share some common problems. Fault displacement hazard refers to the danger of physical movement along a fault plane disrupting the waste emplacement; one can distinguish between principal fault hazard (movement along the fault plane of the earthquake) and distributed fault hazard (secondary movement at some distance from the principal fault plane) (Stepp *et al.*, 2001). Vibratory hazard concerns the possibility of damage due to strong shaking. Secondary hazard due to seismic disruption of ground water patterns is also possible. These topics are covered in a general way in this discussion, which is relevant to a range of geological media. Considerations particularly relevant to argillaceous media are noted where they occur.

The first defence against rupture hazard is to avoid sites that are intersected by potentially dangerous faults, and it is here that categorisation of faults becomes important. Sufficient setback of the waste emplacement from fault planes avoids the obvious problem of principal fault hazard. In the case of faults that are not controlling faults in low seismicity areas, the likelihood of fault rupture is probably rather low: in the UK, for instance, the larger earthquakes typically have depths between 9 and 15 km and rupture dimensions of a few km at the very most; thus the extension of the rupture plane to near-surface is not likely, and has never occurred in historical times.

The question of whether the geological medium (and in particular, argillaceous media) influences the fault displacement hazard, has so far been little studied. It is undoubtedly the case that the nucleation of earthquakes is unlikely within clay or other surficial deposits; according to Marone and Scholz (1988), it is often possible to define an upper stability layer where, apparently, the velocity strengthening character of the upper sediments limits seismogenesis. Counter-examples where a very shallow hypocentral profile is observed tend to be from hard rock intraplate areas without clear faulting, such as the Adirondacks. Even in the United Kingdom, depth profiles show few hypocentres in the upper 5 km, other than small events.

Nevertheless, large enough earthquakes can start in the crystalline basement and rupture through to the surface. The question now has to be posed, whether argillaceous sediments may inhibit this. Evidence that it may do comes from the 1979 Imperial Valley CA earthquake. Quin (1990) showed that a thick upper sedimentary layer, through velocity strengthening, resulted in a zone of negative stress drop above the fault rupture, where rupture propagation was inhibited. Such a process cannot altogether prevent deformation, and a “catching-up” followed the earthquake in which displacement of the sediments occurred aseismically, a process known as afterslip (Crook, 1984).

However, this is a case from a high-seismicity interplate area, and, given the generally higher stress drops from intraplate areas, the effect may be reduced for large earthquakes in lower seismicity areas. Heavy clay deposits in the Mississippi Valley did not prevent rupture propagation either to, or very close to, the surface in the case of the New Madrid earthquakes (Guccione and Marple, 2002). Other examples of faulting to the surface in an intraplate, clay-rich environment could certainly be adduced from the literature.

Assessing the probability of an earthquake occurring of sufficient size to rupture the full width of the crust leads to consideration of the issue of maximum magnitude (Mmax), a fraught topic. For the United Kingdom, one suggestion is that the maximum possible event may be as low as 5.5 Mw (Ambraseys and Jackson, 1985). If this is true, the implications for the possibility of rupture hazard are

significant, and are made explicitly by Ambraseys and Jackson (1985). However, there is no consensus as to how Mmax should be evaluated. Statistical methods (e.g. Kijko and Graham, 1998) tend to produce evaluations not much higher than the historically observed value; arguments from paleoseismicity and "comparable" locations elsewhere in the world can produce higher values but are generally open to debate. There are also disputes as to the magnitude of some key observed historical earthquakes elsewhere, such as the 1356 Basel earthquake.

Turning to the disruption of ground water, it is well known that when earthquake intensity reaches around 7, effects on ground water, such as changes in the flow of springs, become observable (Vogt *et al.*, 1994). Many instances can be advanced even in the relatively low seismicity environment of the United Kingdom, for example the 1884 Colchester earthquake (4.6 ML) (Musson *et al.*, 1990). Fluctuations in the water table can either be short term (due to dynamic response to the passage of earthquake waves) or long term, which may be due to change in pore fluid pressure or an increase in permeability. Expected rises in the water table due to a regional drop in shear stress following an earthquake are of the order of some tens of metres, and are sensitive to the amount of stress change, the compressibility of the whole rock compared to the compressibility of its constituent minerals, and the moisture content above the water table (Cook and Kemeny, 1991). Clay media are likely to have a low ratio of mineral to rock compressibility (K/K_s) and this increases the expected rise in water table.

The issue of shaking hazard is more familiar to seismologists, but its relevance to deep disposal repositories is less clear. Smistad *et al.,* (2001) consider that direct damage from earthquake shaking to the waste packages or engineered barriers of an underground repository is an insignificant chance, but damage to cladding is a possibility and should be evaluated.

Apart from engineering issues, there are two important seismological differences between hazard assessment for a waste repository and for a normal NPP site. The first is the simple fact that a repository is buried at a depth of a few hundred metres. The second is its long operational life.

The depth of burial is significant because strong ground motion decreases with depth. Below 70 m, peak acceleration has been shown to be one-third to one-fifth of its equivalent value at the surface (Sykora and Bastani, 1998). The reduction is greatest for stiff profiles.

The long period of time causes some considerable uncertainties for any seismic hazard assessment. Even when looking at long return period events for normal hazard studies, seismologists are still only concerned with the next 50 years or so of actual time, and can make the assumption that the gross properties of the regional seismicity over that period will be the same as for the last 50 years.

When looking at long periods into the future, this assumption no longer applies, for three possible reasons. Firstly, tectonic processes may change. Just as in the geological past the configuration of plates was different from what it is now, so in the future it will be different again. Many of these developments are more or less predictable from knowledge of current plate velocities; however, significant changes are most likely to take several million years to develop, so this is not really an issue.

Secondly, some tectonic processes have irregularities related to plate geometry. For example, in cases of continental collision, if the impacting plate has a pointed shape, it is likely that two seismic zones will develop on each flank. For mechanical reasons, the brunt of the collision will be taken up by the two flanks alternately rather than both simultaneously. This can result in seismic zones appearing to switch "on" and "off" over periods of several hundred years, and this can be seen with the Arabian Plate and in Iran. If one is dealing with a site that has been "off" over recorded history, it may be difficult to estimate what will happen when it switches "on" again.

Thirdly, climate change can be expected to play a significant role in changing seismicity patterns, at least in the northern hemisphere, a topic that has been the subject of considerable recent research. The possibility of renewed glaciation within the lifetime of a repository has to be considered. There is spectacular evidence from Scandinavia and less clear evidence from Scotland for increased seismicity due to rapid isostatic rebound immediately after deglaciation (Muir Wood, 1989, Firth and Stewart, 2000). Current understanding of the influence of glaciation on seismicity is summed up by Stewart *et al.* (2000). It is expected that the onset of glaciation is liable to suppress earthquake activity, at least for a while, followed by much higher levels of activity after the ice has retreated again, leading in the extreme case to the event (or possibly events) that produced the 150 km long, 13 m high fault scarp of the Pårve Fault in Sweden (Lagerbäck, 1979).

Consequently, estimates of seismicity based on current rates may be quite misleading as to the possible levels of activity that could occur in the more distant future. Modern earthquakes do not much exceed magnitude 4 in those parts of Scandinavia that have evidently seen events in the range 7 to 8 in immediate postglacial times.

Nor is this exclusively a high-latitude problem. Muir Wood (2000) suggests that the large historical earthquakes of the New Madrid sequence (1811-1812) and Charleston (1886) in North America may have been due to an outward moving wave of forebulge collapse from the Laurentian ice sheet, and that a future occurrence of large earthquakes in the presently quiet regions of Georgia, northern Louisiana and Arkansas is perhaps to be expected.

Although it is generally considered that probabilistic methods (PSHA) give a better assessment of seismic hazard for nuclear facilities (Stepp, 2001), some writers are at pains to point out that probabilism and determinism in seismic hazard should not be viewed as competing camps (McGuire, 1999, Bommer, 2002). In particular, McGuire (1999) suggests the principle that, when the worst possible case is reasonably likely to occur, one may as well face up to it and model it deterministically. In other cases, probabilistic methods are a better strategy to combat the problem of over-conservatism.

This principle can be extended by noting that the worst case is unlikely to be observed in cases that are low hazard-short lifetime, but likely to be observed in cases that are high hazard-short lifetime *and* cases that are low hazard-long lifetime.

Also, in cases where conventional PSHA methods are extended to very low occurrence probabilities (such as 10^{-7}), there is a strong danger of obtaining results that may be physically unrealistic, if no account is taken of the physical limits to earthquake ground motion, a subject seldom tackled, and on which there is currently no consensus. Also, even if peak ground accelerations of many g are credible at all, one can reasonably ask what engineering significance they have, given the persistently poor correlation in experience between very high pga values and actual damage to structures.

Whatever procedure one adopts, when working with seismic hazard over long periods of time, inevitably one finds that the critical issues are the limits to the tails of the distributions of parameters – the largest magnitude that can occur; the largest ground motion that can occur for a given magnitude; and the greatest distributed displacement that can occur. Unfortunately, these limits are liable to be poorly constrained; it is the nature of very rare events that we are likely to have little experience of them, even in high seismicity areas. Consequently one must try to walk a fine line between adequate conservatism, and maintaining a sense of physical realism.

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RESEARCH OF FAULT ACTIVITY IN JAPAN

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1. Background

Six hundreds and eighty earthquakes causing significant damage have been recorded since the 7th century in Japan. It is important to recognise faults that will or are expected to be active in future in order to help reduce earthquake damage, estimate earthquake damage insurance and siting of nuclear facilities. Such faults are called "active faults" in Japan, the definition of which is a fault that has moved intermittently for at least several hundred thousand years and is expected to continue to do so in future. Scientific research of active faults has been ongoing since the 1930's. Many results indicated that major earthquakes and fault movements in shallow crustal regions in Japan occurred repeatedly at existing active fault zones during the past. After the 1995 Southern Hyogo Prefecture Earthquake, 98 active fault zones were selected for fundamental survey, with the purpose of efficiently conducting an active fault survey in "Plans for Fundamental Seismic Survey and Observation" by the headquarters for earthquake research promotion, which was attached to the Prime Minister's office of Japan. Forthy two administrative divisions for earthquake disaster prevention have investigated the distribution and history of fault activity of 80 active fault zones. Although earthquake prediction is difficult, the behaviour of major active faults in Japan is being recognised.

Japan Nuclear Cycle Development Institute (JNC) submitted a report titled "H12: Project to Establish the Scientific and Technical Basis for HLW Disposal in Japan" to the Atomic Energy Commission (AEC) of Japan for official review [1]. The Guidelines, which were defined by AEC, require the H12 Project to confirm the basic technical feasibility of safe HLW disposal in Japan. In this report the important issues relating to fault activity were described that are to understand the characteristics of current fault movements and the spatial extent and magnitude of the effects caused by these movements, and to estimate the spatial extent of future fault movements and their effects on the geological environment. One conclusion from the report is that present active faults in Japan have moved repeatedly for at least the last several hundred thousand years and are likely to continue to do so under the same stress field in the next a hundred thousand years. The latest knowledge relating to active faults and features of active faults to be considered for the stability of geological environments are described.

2. Characteristics of fault activity in Japan

Tectonics of the Japan islands and faults activity

The Japanese islands and the adjacent sea areas are located near the margins of the Pacific Ocean, Philippine Sea, Eurasian and North American plates (Figure 1). The Pacific Ocean plate and the Philippine Sea plate subduct beneath the continental plates on the west side. The current general plate

movements were established by the end of the Miocene to the end of the Pliocene [2]. Consequently, when estimating the spatial distribution of natural phenomena in Japan for the next hundred thousand years, it is reasonable to consider that the patterns and trends in tectonic activity over the past several hundred thousand years will continue. It is also justifiable to assume that geological phenomena as associated with the tectonic setting of Japan will progress with predictable trends for at least the next hundred thousand years. Figure 2 shows the result of calculating the deformation rate and its orientation in the Japanese archipelago based on the measurement value obtained by Global Positioning System (GPS). It has been shown that strong compression is now occurring in an east-west direction underneath the Japanese archipelago. A similar conclusion has been obtained from an analysis of seismic observation data. The results of various topographical and geological surveys form the basis for the hypothesis that a similar compression force has been exerted underground for quite some time, at least for several hundred thousand years [3].

Figure 1. Plates around Japan

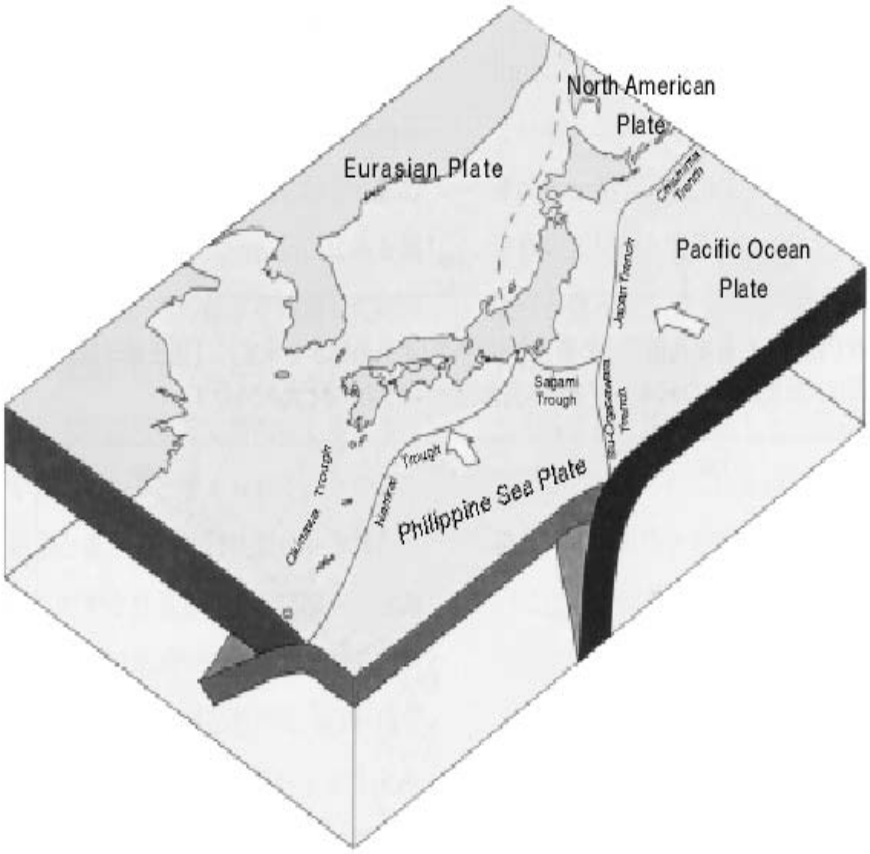
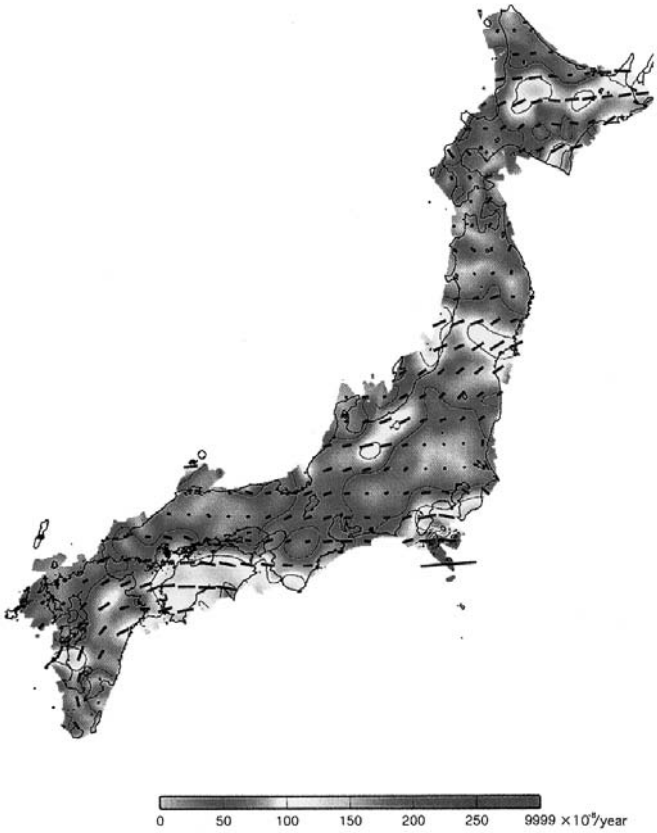


Figure 2. Crustal deformation



The history of fault activity, such as displacement, latest active time and recurrence interval, has been estimated from investigation of the development and frequency of occurrence of tectonic landforms and key beds (e.g. tephra), often by observation of outcrop. We compiled the results of relevant previous investigations. In Japan, most active faults are located in a compressive stress field caused mainly by the relative movements of the tectonic plates. For example, northeast Japan has been located within an east-west compressive stress field for several million years and many reverse faults have developed.

The situation in southwest Japan is slightly more complex [4], but the current fault movement was maintained across much of the area by 0.5 million years ago. No significant changes of orientations and of the rates of displacement have been observed in major active faults in the past several hundred thousand years. Thus, present major active faults have been active for at least several hundred thousand years and are likely to remain active.

Distribution and types of active faults in new active faults map

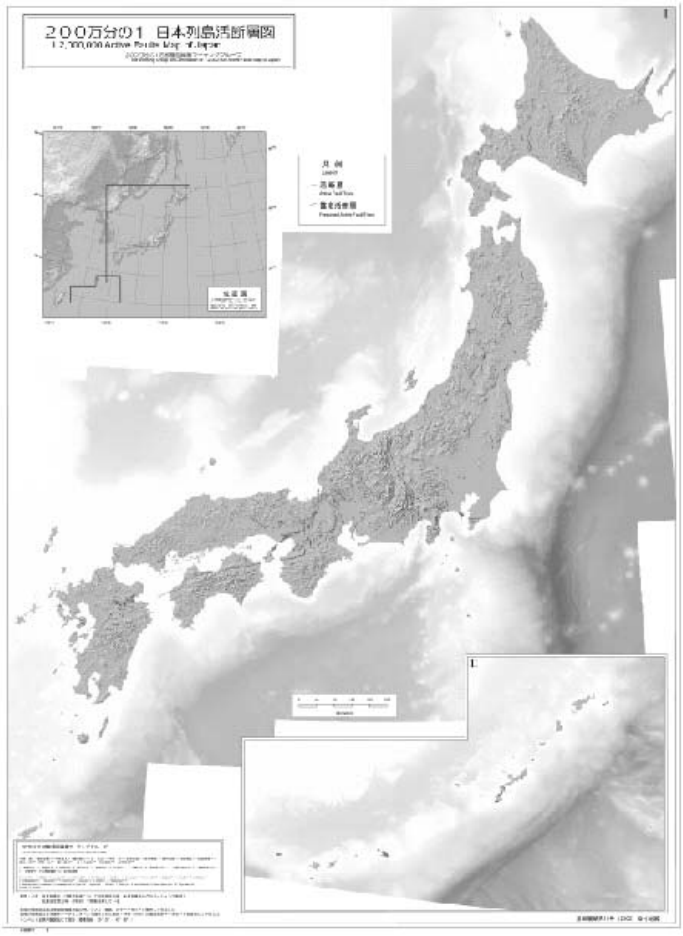
We compiled new active faults map of Japan (Figure 3) based on the digitised Geographic Information System (GIS) data of active faults with a scale of 1:25 000 [5]. Active faults were defined in the new map as faults that have repeatedly moved during the late Quaternary with intervals of one thousand to ten thousand years forming distinctive fault-related tectonic features on the earth's surface. Previous studies [6] commonly defined active faults as faults that repeatedly moved during

the Quaternary and are the potential source of future earthquakes. Broadly speaking, the localisation of active faults that have prevailed for the last several hundred thousand years can be seen throughout the Quaternary period. The polarisation of the distribution of active faults and the localisation of their degree of activity and types are also described in previous studies. Active faults clearly show localism of distribution densities and types, which reflect differences in geological structures and physical properties of the upper crust. Reverse faults with a north- south strike are concentrated in northeast Honshu and in the Kinki area. In contrast, left-lateral faults with a northeast-southwest strike are concentrated in the west of central Japan. The Izu Peninsula and its northern adjacent district differ from the other districts in trends and types of active faults.

Extent of the range of active fault movement and active fault zone

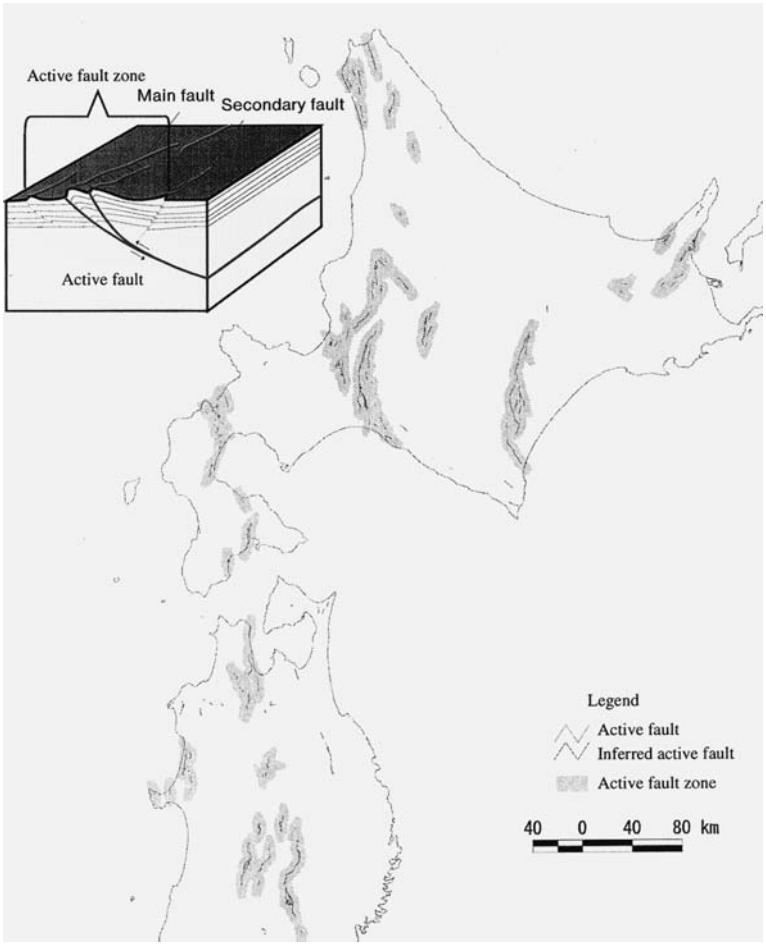
An evaluation was made of whether or not a reverse fault zone, which consists of several reverse faults running nearly parallel to one another, has extended. Geophysical exploration was conducted across a reverse fault zone at the eastern edge of the Yokote basin. This study showed that, during the past several hundred thousand years, the fault zone has propagated over a range of about 2 km near the surface. All findings indicate that possible future propagation of any reverse fault zone is unlikely to exceed 8 km and occurs in the direction of the basin.

Figure 3. New active fault map of Japan



Many large earthquakes in the past were often caused by the movement of several active faults that distribute in the vicinity of each other. Matsuda [7] grouped a series of faults, designating them as "seismogenic faults" with the potential to independently cause earthquakes. Such a grouping of faults is important for estimating the behaviour of individual active faults. On the basis of the latest map of active faults [5], the division of active fault zones exemplifies the fields of future faulting (Figure 4). A fault zone is defined by the envelope surrounding individual active faults that have: i) similar orientations; ii) separations parallel to the fault planes up to 5 km; or iii) separations perpendicular to the fault planes of up to 10 km. By comparing the overall properties of such a fault zone with the properties of individual active faults within the zone, it should be possible to infer general properties of faults within the zone that have not been identified explicitly. The identification of active fault zones could lead to a broad understanding of the extent of their expansion in future. There have been no cases of new active fault zones appearing at least during the last several hundred thousand years. The geological information mentioned above indicates that, in an area where significant changes in the crustal stress field are unlikely, there is little possibility that new fault zones might be generated independently of movements along existing active fault zones.

Figure 4. Example of active fault zone



3. Effects of fault movement

The mechanical influence of fault activity on the geological environment is limited mainly to fault crushed zones. Otsuki [8] points out that field and laboratory data on faults, both in Japan and abroad, indicate a relationship between the width of a crushed zone (mainly fault gouge and fault breccia) and the cumulative displacement. This relationship can be used to estimate the widths of fault-crushed zones from the scale (length and cumulative displacement) of the associated faults. Based on this relationship, the crushed zone of a fault with a cumulative displacement of 10 km would be roughly 200 m in width. The extent of this influence is several meters to several hundred meters at most. At the time of the 1930 Kita-Izu Earthquake, a displacement of about 2 m was observed in the fault-crushed zone in the Tanna Fault (length: 30 km). Deformation was observed to have occurred in an area extending for about 1.5 km to the west of the fault crushed zone [9]. The area where fractures are distributed in a rock mass formed by fault movement, is reported to be in the order of 10^{-2} relative to the length of the fault [10]. The range over which associated small faults are distributed in the surrounding rock mass is considered to be within a few kilometres of large active faults.

The location and scale of potentially perturbing faulting have been traced back over the past several hundred thousand years. Based on the results of these studies, it can be shown that it is possible to select a sufficiently stable geological environment in Japan, such that a repository will not be influenced by disruptive faulting for at least the next hundred thousands years. A detailed investigation of the closest active faults to any future site proposed for HLW disposal should be performed. This investigation should allow an appropriate distance between the disposal facilities and active faults to be defined. This is necessary because the size of regions affected by fault activity depend on the sizes of the active faults. In addition, an in-depth investigation should be performed at the site to identify whether or not there are small active faults that were not recognised during regional surveys. This approach will avoid the influence of such faults during detailed siting. The effects of earthquakes also should be considered during the design of the geological disposal system. However, the influence of seismic vibrations several hundred meters underground or deeper is generally regarded to be small.

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GEOLOGICAL EVOLUTION OF CLAY SEDIMENTS: THE PETROLEUM EXPLORATION VISION

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Introduction

The radioactive waste isolation capacity assessment for a clay sediment host rock is link: (1) to the understanding of their present state properties and 3-D repartition (from basin evolution, including sedimentary and diagenetic process); and (2) to the prediction of their future evolution during the next million years.

For petroleum exploration, basin modelling aims at reconstructing the accumulation of hydrocarbons at basin scale, and at geological timescale, taking into account the effects of kinematics displacements, sedimentation, erosion, compaction, temperatures history, overpressures and fluids flows (water and hydrocarbons). Furthermore, explorationists wish to address overpressure reconstruction in order to estimate the risks of drilling.

Clay sediments are of interest for petroleum exploration because source rocks and seal are generally composed of them. Nevertheless, in spite of their occurrence in nature their evolution at geological timescale is not well understood. And, most of the knowledge has been achieved by those working in the realms of soils mechanics and civil engineering until the present geological investigations for long term radioactive waste repositories. Application of this knowledge to clay sediment is considered to be valid within the first hundreds of meters at the top of the sedimentary pile, according to a repository depth.

This paper is dedicated to the sedimentary rocks behaviour at geological timescale. This behaviour is characterised by: (1) the deposition of the sediment; (2) the loading path at geological timescale; (3) the constitutive law which includes the consolidation process and the rupture criteria; and (4) the parameters evolution related to consolidation.

Deposition

Clay sediments are generally deposited in low energy environments. Geometry, composition and texture of the bodies are also controlled by the depositional environment. More generally, sediments architecture is the response to complex interaction between three processes: (1) the creation of available space of accommodation in the basin; (2) the supply of sediment brought into or produced in the basin; and (3) transport of sediment within this basin (e.g. Grangeon, 2002). These processes can now be simulated with dynamic-slope forward computer model such as Dionisos (Granjeon & Joseph, 1999). This kind of model is able to simulate the stratigraphy in 3-D on a prospect and on a basin-wide scale. And, if they are often used to predict the distribution of the reservoir bodies, their results can be

interpreted in term of distribution of muds. Furthermore, the intercalation of turbidites or other geological bodies such as sedimentary sills or dykes (e.g. Parize *et al.*, in prep.) can modify drastically the permeability tensor: in some case the macroscopic permeability along the stratification may be up to five order of magnitude greater than the permeability across the stratification.

Loading path

One key point for simulating the behaviour of sediments at basin scale is to be able to compute the stress tensor. This is generally not addressed in basin model because of the complexity of this problem. (Luo *et al.*, 1998). In order to estimate the stress tensor all along the geological history, it is generally assumed that the total overburden weight (σ_v) is one of the main stresses. The second assumption is that the two other main stresses are horizontal and equal ($\sigma_H=\sigma_h$). Furthermore, we assume that the total horizontal stress can be deduced from the total vertical stress by the following formula: $\sigma_h=Ko(z).\sigma_v$. According to the tectonical environment, Ko may varies from values lower than 1 in extensional area to values greater than 1 in compressive area (e.g. Grauls, 1996).

Constitutive laws

From the top to the bottom of the sedimentary column we distinguish four major horizons. The first horizon at the top of the sedimentary column is made up of mud. Beneath this horizon we consider that the sediment forms a soil having a thickness of some meters. The appearance of cemented grains marks the start of the third horizon. It is made of the sedimentary rock *sensu stricto*. The beginning of the metamorphism marks the end of this third horizon and the beginning of the fourth one.

The previous geometrical description of the sedimentary column could be considered as an historical evolution of sediments that are mud when they are young, then they become soils, sedimentary rocks and metamorphic rocks. Determining all the mechanisms that cause this evolution would appear to be impossible. However, attempting to determine the major physical phenomena in the history of the sediment is feasible. Hedberg (1936) has already tackled this operation. Here we will only consider the sedimentary rock *sensu stricto*, and will use the word consolidation to represent all the physical processes that act on the sediment during its geological history.

Figure 1. Porosity effective stress curves for natural and lab-compacted

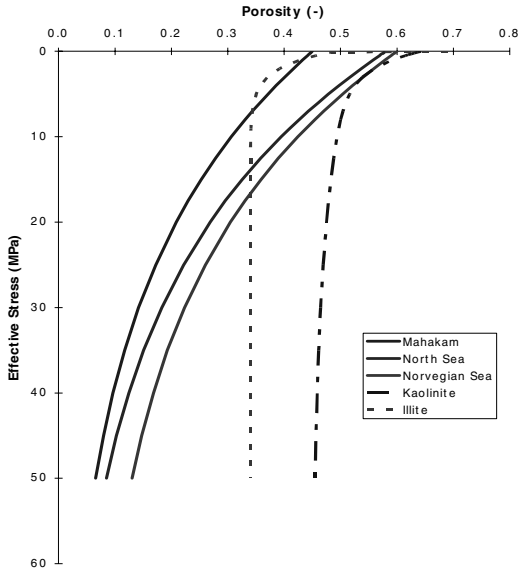
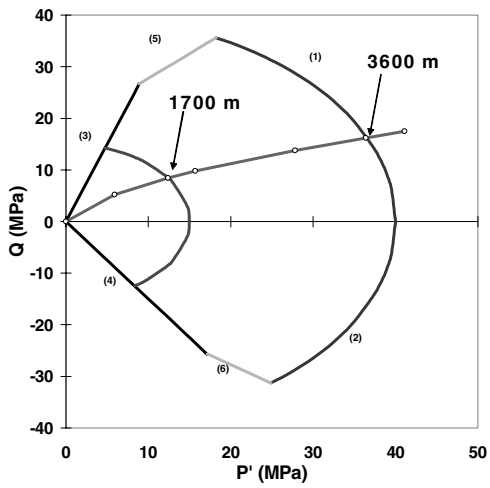


Figure 2. Example of normal hydrostatic stress path ans corresponding elastoplastic yield



Consolidation is considered here as the combined effect of the mechanical compaction and the chemical compaction (Schneider *et al.*, 1996). The mechanical compaction is mainly caused by the rearrangement of grains during burial. This mechanism is efficient near the surface. The mechanical compaction could be represented at the macroscopical scale by an elasto-plastic rheology. The chemical compaction is considered here as resulting from dissolution-precipitation mechanisms, generally induced by stress (pressure-solution). This mechanism generally becomes to be active at a depth varying from few hundred to few thousands meters. The chemical compaction could be represented at the macroscopical scale by a visco-plastic rheology. The resulting elasto-visco-plastic rheology could be written as follows:

$$\frac{d\phi}{dt} = -\beta(P')\frac{dP'}{dt} - \alpha(\phi, T)P'$$

The visco-plastic rheology has been introduced to account for pressure solution in limestones and sandstones. Nevertheless it is also used for shales because a viscous component exists as shown in Figure 1 where porosity lower than 30% cannot be reached by mechanical compaction only.

At a given time, the consolidation elastoplastic yield is supposed to be given in the (Q, P') space by the classical following equation:

$$Q^2 + P'^2 - P'^2_c = 0$$

where P'_c is the consolidation mean effective stress, which evolves with the geological history of the sediment. P'_c is not only a function of the maximum mean effective stress reached by the sediment, as assumed in soils mechanics, but also a function of time and temperature due to chemical reactions. This is what we call the chemical hardening of the sediments.

The complete elastoplastic yield (Figure 2) is defined by the union of the consolidation elastoplastic yield and of the different failure criteria that could be seen as elastobrittle yields (Schneider *et al.*, 1999). Thus, the elastoplastic yield is composed of six elementary elastoplastic yields which define the onset of: (1) vertical compaction; (2) horizontal compaction; (3) vertical tensile fracturing; (4) horizontal tensile fracturing; (5) subvertical shear fracturing; (6) subhorizontal shear fracturing.

Parameters evolution with consolidation

Due to the consolidation, most of the parameters that describe the physical properties of the sediments evolve with the geological times. One difficulty is to quantify the degree of evolution of the porous medium during its geological history. We have chosen to measure the evolution of the sediments by their porosity because this is one of the physical properties that can be measured easily. This approach is only valid when there is no strong diagenesis with creation of high secondary porosity.

Mechanical properties

The elasto-plastic coefficient is given by (Schneider *et al.*, 1996):

$$\beta(P') = \frac{\phi_a}{E_a} e^{-P'/E_a} + \frac{\phi_b}{E_b} e^{-P'/E_b}$$

where ϕ_a, E_a, ϕ_b, E_b are coefficients that characterise the sediment. These coefficients may be calibrated by using observed porosity-depth data from sediment that have not suffer diagenesis or using experimental laboratory measurements (Schneider *et al.*, 1996).

The visco-plastic coefficient is given by (Schneider *et al.*, 1996):

$$\alpha(\phi,T) = (1-\phi)/\mu_s^o e^{\frac{E}{R}\left(\frac{1}{T}-\frac{1}{To}\right)}$$

where μ_s^o is the macroscopical sediment viscosity at T^o, and E is the activation energy. These coefficients may be calibrated from set of observed data (Schneider *et al.*, 1996; Schneider & Hay, 2001).

The Young's modulus and The Poisson's coefficient could be expressed as a function of the porosity. For example, for chalk, Engstrom (1992) proposed the following relations:

$$E = 224.8\, e^{-11.2\,\phi} \quad \text{(GPa)} \quad \text{and} \quad \nu = 0.117 + 0.21\,\phi$$

In the plastic domain we suppose that the effective stress coefficient is equal to 1. In the elastic domain, the Biot coefficient could be expressed as a function of the porosity (Schneider *et al.*, 1993):

$$b(\phi) = \frac{\phi}{\phi + a(\phi^0 - \phi)}$$

The tensile strength (T) of sediment is generally weak (e.g. Lama & Vutukuri, 1978). If we assume that the sediment is fractured, the tensile strength could decrease to 0. Indeed, the criterion for the opening of existing fractures is the same we wrote for a tensile fracture but with T = 0.

The cohesion and the friction angle of the sediment are generally known to increase with depth or to increase as the porosity decreases. They could be expressed as functions of the porosity and mineral content. For example, for chalk we can use the empirical relations (Rhett *et al.*, 1992):

$$c = 20.9 - 41.1\,\phi \quad \text{(MPa)} \quad \text{and} \quad \varphi = 56.1 - 105.6\,\phi \quad (°)$$

Once the expression of the coefficients as functions of the porosity are given, the evolution of the different elastoplastic yield surfaces can be drawn in the (Q, P') space. This has been done in Figure 2 in which the yield surfaces have been drawn for chalk at 1.7 km and 3.6 km. In this figure, it should be noticed that at 1.7 km depth, the elastic domain is defined by the tensile failure criteria and the consolidation yield surface while the elastic domain needs all the rupture criteria to be defined at 3.6 km.

Hydraulic properties

Sediment permeability is given as the sum of two terms. The first one which only depends on the porosity represents the matrix porosity. The second one which is both a function of the pore pressure and the porosity represents the fracture permeability. The matrix permeability is given by an empirical formula, which is an extension of the Koseny Carman formula: $K_m(\phi) = 0.2\phi^n / S^2(1-\phi)^2$. The general permeability model, which includes the fracture permeability is given, when Ko is lower than one, by the following empirical formula:

$$p \le \sigma_h \quad K_f(\phi, p) = K_m(\phi)$$
$$p > \sigma_h \quad K_f(\phi, p) = K_m(\phi)\left(1 + \alpha_f\left(\frac{p}{\sigma_v} - Ko\right)^{a_f}\right)$$

Conclusions

Prediction of the behaviour of clay sediments in the future needs to reach a good knowledge of their present state. On one hand, the geometry, the texture and the composition of clays sediments bodies are function of the geological environment during deposition. On the other hand, the present

day physical properties of sediments are the result of a geological history characterised by a loading history (P, σ, T) and related mechanical and chemical modifications.

One of the properties of the sediment is its elastoplastic yield, which evolution is characterised by a long term chemical hardening. The relative position of the stress state at present day which respect to the elastoplastic yield is crucial for any forecasting. For example, the impact of a pore pressure modification will be different in a basin where sedimentation is active (the stress tensor is initially on the elastoplastic yield) and in a basin where erosion is active (the stress tensor is initially in the elastic domain).

If a sediment can be characterised through lab experiments, large timescale processes can only be addressed by basin modelling. Nevertheless, basin simulators needs physical models that are able to reproduce the evolution of clays sediments through geological times. These models are built accordingly to our understanding of the diagenetic evolution of shally formations.

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PREDICTING THE DIAGENETIC EVOLUTION OF ARGILLITE REPOSITORIES BASED ON THE STUDY OF NATURAL AND EXPERIMENTAL SYSTEMS

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1. Introduction

Understanding the nature of diagenetic reactions involving clays and associated mineral phases is an important aspect in predicting the long-term chemical and physical properties of host-rock argillite in underground radioactive repositories. Much of our current knowledge is based on the study of mineral phases in natural rock sequences that have formed under a range of pressure (P), temperature (T), compositional (X) and time (t) conditions (Velde, 1992). Valuable information has also been obtained by reproducing reactions in the laboratory under constrained conditions using batch or flow-through reactors (e.g. Nagy, 1995; Wogelius & Vaughan, 2000). Based on our current knowledge, six primary controls can be recognised, which drive diagenetic reactions in the upper portion of the Earth's crust.

i) Fluid chemistry, providing the essential ingredients for exchange of cations and anions.

ii) Rock composition, laying down the basic building blocks and reaction sites.

iii) Rock permeability, necessary for allowing material transport and fluid-rock interaction.

iv) Temperature, as the principle energy source for enhancing reaction rates.

v) Time, needed to allow reactions to get closer to their equilibrium conditions.

vi) Stress and strain, for mechanical mixing of reactants and products.

In order to predict the diagenetic evolution of argillite repositories, we need to know firstly, which reactions occur under a defined set of conditions, secondly, how these reactions modify the material properties of the argillite seal, and finally, how fast these chemical reactions take place. Based on the application of thermodynamics, and the construction of activity diagrams for low temperature mineral phases (e.g. Velde, 1992), fair predictions of mineral stability can be made under a given set of physical and chemical conditions. Such predictions are strengthened by examining mineral reactions that occur under natural conditions and also those that have been preserved in the geological record. Changes in the material behavior can also be reasonably assessed, as the basic physical and chemical properties of argillaceous rocks of varying mineralogy are well documented in the petrophysical and engineering literature (e.g. Bell, 1999).

One of the most difficult tasks in predicting the response of argillites to changing physical and chemical conditions is to assess the rates of the chemical reactions involved. This difficulty reflects

our poor knowledge of the reaction kinetics for these low-temperature, fine-grained mineral materials. The aim of this contribution is therefore to review the timing and duration of diagenetic reactions that occur in nature and to compare them with reaction rates measured by laboratory experimentation. Some new analytical techniques are also presented, which should serve as a better analogue for investigating and predicting the outcome of radioactive repository sites under the influence of varying geological scenarios.

2. Diagenesis in the geological record

The geological record is unique in that it provides a natural laboratory that has been running since the formation of the Earth. The rock record extends back to ca. 3.9 Ga, and there are case studies of ca. 2 Ga old illitisation of argillites in contact with radioactive bodies (Bros *et al.,* 1992). Traditionally, clay mineral diagenesis of argillites has been considered to be largely a function of temperature and time (Weaver, 1960). Generally, older sedimentary sequences are often more mature or metamorphosed, with progressive mineral alteration and recrystallisation occurring in response to successive burial within the Earth's crust (Clauer & Chaudhuri, 1995). Currently, rock alteration and diagenetic reactions are more favorably viewed as occurring during discrete geological events, in response to distinct changes in physico-chemical conditions. Even the classic Gulf coast diagenetic sequence of Hower et al (1976) has been reinterpreted as resulting from number of punctuated events (Eberl, 1993). The timing of such events can be dated isotopically. For example, Rousset & Clauer (2003) determined the age of diagenesis in the Callovo-Oxfordian argillites within the Paris Basin to occur 10-15 Ma after deposition of the sediments, related to a decline in sea-level. In the case of mixed illite assemblages, the isotopic ages of the various growth events may be modeled, as by Srodon *et al.,* (2002).

The rates and duration of diagenesis are much more difficult to assess from the geological record. Whereas the rate of mineral dissolution during surface weathering can be measured (Schnoor, 1990), the kinetics of dissolution and crystal growth reactions at depth in the Earth's crust can only be extrapolated. There are, however, a number of situations where the rates of diagenesis can be approximated from the geological record. The intrusion of magmatic material into argillites provides a natural analogue for underground nuclear waste repositories, where the nature of thermally induced mineral reactions can be characterised. Techer *et al.,* (in press), for example, document the formation of illite-smectite mixed-layered clays, and the disappearance of kaolinite and chlorite, toward the contact of a 2 Ma old, 1metre thick dike. These reactions appear to occur rapidly at temperatures of <260°C and probably within a duration of a few to tens of years. The rapid growth of clay minerals is also evident from studies of seismically active fault zones, such as the San Andreas Fault System of California and the Alpine Fault of New Zealand (Chester *et al.,* 1993; Warr & Cox, 2001). Here, pore-filling low temperature mineral phases, such as clays, zeolites and carbonates, are intimately linked with the earthquake cycle and form during the inter-seismic periods that separate high velocity slip events.

3. Simulating diagenesis in the laboratory

Laboratory experiments have the distinct advantage that P-T-X conditions can be carefully controlled and measured, but the disadvantage that only short reaction times can be studied. Most quantitative studies of reaction kinetics experiments have focused on the study of mineral dissolution rather than crystal growth, and changes in mineral abundance are typically measured indirectly by monitoring changes in solution composition. Reviews of the available experimental data have been compiled by Nagy (1995) and Kloprogge *et al.* (1999).

3.1 Batch reactors

Batch reactors are the most commonly used form of reactant vessel for simulating mineral alteration and diagenesis. These reactors are closed systems, and in order to facilitate rates, a large water to rock ratio is used, along with occasional stirring to mix reactants and products.

Experiments relevant to the diagenesis of argillites under various repository conditions have been conducted. Rapid alteration of smectite has been shown to occur under low temperatures (< 60°C) and alkaline solutions (pH > 13) caused by the alteration of concrete (e.g. Rassineux *et al.,* 2001). Dissolution of smectite and formation of a random illite-smectite phase was observed in similar experiments on Callovo-Oxfordian argillites from the Meuse-Haute Marne laboratory (Claret *et al.,* 2002). However, here the organic coating on particle edges appears to reduce the reactivity of the clay minerals.

Rates for the conversion of smectite to illite at higher temperatures (250 to 350°C) have also been studied experimentally for varying K+ in solution (Huang *et al.*, 1993). Based on these experiments, a kinetic equation was presented, which can be used to predict the extent of reactions in sedimentary brines largely dependent on time, temperature and potassium concentration. The effectiveness of this, and other available kinetic models, has been evaluated by Elliot & Matisoff (1996) for various geological settings.

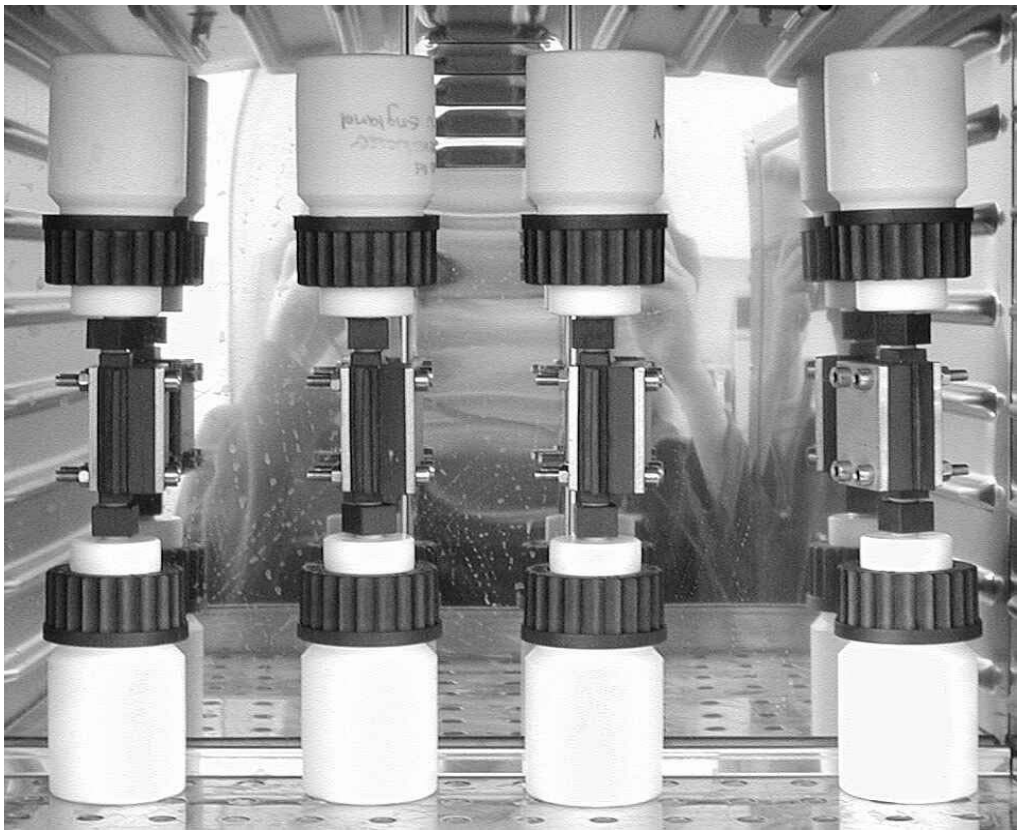
3.2 Flow-through reaction chambers and wet-cell X-ray diffraction techniques

Flow-through reactors are not commonly employed for the synthesis of low temperature mineral phases in the laboratory, probably because of their complexity. They do, however, provide better analogues for the Earth's crust, with a dynamic open system characterised by the circulation of fluids and mass transfer.

A new approach to monitoring the reaction kinetics of fine-grained minerals in solution has been presented using wet-cell X-ray diffractometry (Warr & Hofmann, 2003). This device consists of a flow-through reaction chamber, which can be routinely mounted onto the X-ray diffractometer for in-situ measurements of the sample. With the aid of a pressure cap, the device can be used under hydrothermal conditions (<150°C). Experiments are currently running in order to quantify the rates of diagenetic reactions in argillite host rocks under repository relevant conditions.

The setup in Figure 1 shows cells containing pressed powders from the Callovo-Oxfordian Formation of the Meuse-Haute Marne underground repository site. The reactive solution passes from the upper source bottle, percolates slowly through the cell, and collects in a lower catchment bottle. At 85°C the solution is driven largely by differences in vapour pressure. X-ray diffraction (XRD) data is obtained at various time intervals with minimum disturbance of the sample by measurement through a caption foil for the duration of the experiment. Time-dependent changes in the intensity of XRD reflections are used to monitor changing abundance of reactants and products and to establish kinetic relationships. The composition of the circulating solution can be monitored, and the role of permeability on reaction rates assessed.

Figure 1. Flow-through reactors (wet-cells) been used to investigate the kinetics of diagenetic reactions in Callovo-Oxfordian claystones from the HTM102 core of the Paris Basin
Reaction temperature = 85°C, solution = simple young fluid (Claret *et al.,* 2002)



4. Conclusions

Currently we know a lot more about which diagenetic reactions should occur in argillites under a fixed set of P-T-X conditions, but our knowledge of reaction rates remains limited. Therefore we can predict the type of reactions that occur in response to a particular geological scenario, but quantifying speed and duration is still of a speculative nature. Better assessment of the kinetics of diagenetic reactions is required under repository relevant conditions through the study of experimental and natural analogues.

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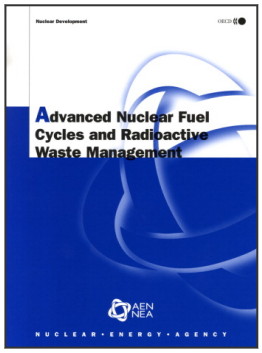

From:

Stability and Buffering Capacity of the Geosphere for Long-term Isolation of Radioactive Waste

Application to Argillaceous Media - "Clay Club" Workshop Proceedings - Braunschweig, Germany, 9-11 December 2003

Access the complete publication at:

https://doi.org/10.1787/9789264009097-en

Please cite this chapter as:

OECD/Nuclear Energy Agency (2005), "Exogenic and Endogenic Processes", in *Stability and Buffering Capacity of the Geosphere for Long-term Isolation of Radioactive Waste: Application to Argillaceous Media - "Clay Club" Workshop Proceedings - Braunschweig, Germany, 9-11 December 2003*, OECD Publishing, Paris.

DOI: https://doi.org/10.1787/9789264009097-4-en

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