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UPLANDS AND GLOBAL ZONATION

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Dedication: This paper is dedicated to the four symposium honorees on the occasions of their respective milestone birthdays: Prof. emerita Kazue Fujiwara (70), Prof. emeritus Franco Pedrotti (80), Prof. Vasile Cristea (65), and Dr. Gheorghe Coldea (75).

Abstract: Attempts to regionalize world climates began in the early 1800s, based mainly on empirical relationships with natural landscapes. Eventually, though, increasing awareness of global atmospheric circulation yielded "genetic" classifications and zonation concepts, based on atmospheric mechanisms. It was generally accepted that zonation follows temperature patterns, with wet and dry seasons as complicating factors. All concepts, however, were based on patterns in lowland areas – but what are lowlands? And how might bioclimatic zonation be extended systematically to upland areas?

Uplands are usually recognized as areas where natural landscapes are significantly different from those in lowlands. But uplands can also be delimited by measurable criteria, such as topographic "break" points; the onset of limiting temperatures or steeper temperature gradients; or by thresholds or steep gradients of improving water balance, which may involve air humidity as well as precipitation or decreasing severity of a dry season. On islands and along coastlines, atmospheric inversion layers and persistent clouds may also delimit uplands.

The effect of elevation on dry seasons, in particular, is projected herein for 1 km and 2 km upshifts in those climate types that have dry periods. Suggestions are also made for temperature and water-balance criteria to identify upland onset and higher altitudinal belts, which are affected by temperature lapse rates and vertical precipitation patterns that are not consistent geographically, seasonally or on different slope aspects. Finally, the magnitude of temperature inversions is also demonstrated, but how to estimate and include them in broad-scale geographic models remains illusive.

Keywords: altitudinal belt, bioclimatic zonation, climatic water balance, dry-climate amelioration, expanded Walter climates, humidity, seasonal lapse rate, temperature inversion, upshift effect, vertical precipitation pattern

Introduction

Attempts to regionalize world climates and vegetation began in the early 1800s, based largely on recognition of differences in vegetation and natural landscapes. The first approaches were empirical, but increasing awareness of global atmospheric circulation eventually yielded "genetic" classifications and zonation concepts, based on atmospheric mechanisms and temperature patterns. Wet and dry regions were more complicated, but it was generally clear that zonation should follow from temperature patterns. Various summaries of what can be called bioclimatic zonation have been presented, most recently by Box [14], but all such systems are based on conditions in lowlands.

So what about upland zonation? The concepts of geographic 'zones' (horizontal) and altitudinal 'belts' (vertical; *Stufen*, *étages*) were formalized at the International Botanical Congress of 1910. The terms 'alpine' (meaning above treeline), 'colline' (in foothills), and 'montane' were also defined, the latter though a bit differently from now [78]. These terms came quickly into standard usage, with 'subalpine' (just below treeline) replacing 'montane' and moving it downward to apply to the main vertical extent of most mountain systems [e.g. 84, 111, 17, 31]. Fundamental to these zonal concepts, though, was tacit agreement that they are all abstractions and are recognizable only by their [concrete] vegetation types. For example, Braun-Blanquet (1951) defined altitudinal belts as 'bioclimatic vegetation belts'. Even so, efforts were also underway to study and understand how climatic conditions change in mountains and how these changes relate to vegetation changes [e.g. 97]. Global classifications were developed but were based always on relationships in "lowlands". For example, in the global climate system of Köppen (1931) all mountain areas are grouped as climate H (for highlands); and in the system of Walter (1970, 1976) [cf. 109], mountain climates are listed after the nine main types as type X.

Lowlands are usually recognized, quite empirically, as those regions below the point where the natural landscape changes significantly to something else. Uplands are thus construed as areas with different vegetation, such as mixed or purely coniferous forest above areas of deciduous broad-leaved forest. But uplands can also be recognized by topographic criteria such as escarpments or other "break" points, where elevation begins to increase more rapidly. And uplands are usually cooler, more humid and have more precipitation (up to cloud level). So uplands may be recognizable by climatic criteria, such as the onset of limiting temperatures or steeper temperature gradients; or by thresholds or steep gradients of improving water balance. On islands (e.g. Tenerife) or along coastlines with cold offshore currents (e.g. Peru), atmospheric inversion layers and persistent clouds may also serve to delimit what may effectively constitute uplands.

The purpose of this paper, then, is to consider where uplands start and how to include uplands, and the criteria needed to distinguish them, in complete global zonation systems. What are these criteria and measures, which are not constant in all parts of the world or even at all times of the year – or even on all sides of a given mountain? Quantitative values for the delimitation of uplands must include both means and minima for growing-season and winter temperatures. Quantification of water-balance criteria is more difficult, but some temperature and water-balance criteria are suggested. In order to place uplands, though, we must start with a global geographic framework of lowland climate types.

Global Framework

It was recognized early that the basic zonal divisions of the Earth depend primarily on temperature and that zonal patterns are related to patterns of global atmospheric circulation [see 14]. Global climate classifications thus gradually evolved toward this 'genetic' basic, i.e. a basis in the mechanisms that generate the different climatic types and regions. These mechanisms include global circulation but also latitude (through its effect on solar input) and what is called 'geographic position', i.e. the relative position of a place on a land mass (e.g. windward, interior, leeward-coastal).

Advantages of genetic climatic classification systems include the following:

- Types are tied more directly to mechanisms that generate different types, i.e. to patterns of global atmospheric circulation and geographic position
- There are fewer empirical estimates of limiting temperatures or precipitation ratios
- Boundaries and seasonal durations may be tied to limitation mechanisms that have biological significance and are thus expressed in natural vegetation
- Mechanism-based systems are more flexible under change than are empirical systems, which have limits that may become invalid with changing global relationships.

The most complete, best known of the various genetic classification systems is that by Heinrich Walter (1954); [cf 37, 6], with its set of 9 basic climate types and Roman-numeral notation:

I. Equatorial	VI. Typical Temperate
II. Tropical Summer-Rain	VII. Temperate Continental (and drier)
III. Subtropical Arid	VIII. Boreal
IV. Mediterranean	IX. Polar
V. Warm-Temperate	

This system did, however, have a (very) few shortcomings. It did not distinguish east-side versus west-side types called 'warm-temperate' (type V), which are caused by quite different atmospheric mechanisms, have quite different general temperature levels, and hardly even occupy the same latitudes. The Walter system also has only 9 types, with no explicit system of subtypes. Finally, it has no system for differentiating mountain climates – but neither do other systems.

A solution was presented by Box [14; cf 13], as follows:

- Split Walter's V into a **Marine West-Coast** (Vm) type and a **Warm-Temperate** (Ve) type, the latter occurring mainly on continental east sides and at much lower latitudes;
- Improve global consistency by tying some limits to known biotic limitation mechanisms;
- Add necessary **Subtypes**, and recognize **Transitions** more explicitly; and
- Recognize explicitly and explain **Exceptions** to the regular zonation pattern.

The result can be called the 'enhanced Walter system' and has also been displayed graphically on an "Ideal Continent" [13; 16]. Naturally occurring zonal soil types, corresponding to zonal climates, were recognized by Dokuchaev [29, 30], and corresponding zonal (i.e. climatic-climax) vegetation types were presented already in the first version of Walter's system [103]. The enhanced Walter system, with zonal vegetation types, is shown in Table 1 and serves as a consistent global basis for classifying climates and natural vegetation regions, and as a basis for considering vertical zonation in mountains. Walter himself began this task of mountain zonation by introducing the concept of orobiome [107; cf 108]. The zonal vegetation for each altitudinal belt in each zone also appears to be consistent (edaphic problems aside, see vegetation descriptions by Walter 1968, 1973) – but more about this later.

A listing of the world's main upland areas is shown in Table 2. For each upland area there is some indication of its geographic shape (e.g. linear versus plateau) and relationship to climate zones (single or multi-zonal, in the enhanced Walter system). Uplands that occur entirely within one climate zone can be treated simply by adding the usual altitudinal belts as a third dimension. Multi-zonal mountain ranges, such as the Urals, are more complicated [cf 107].

Table 1: Mechanism-Based World Climate Zones and Zonal Vegetation Types

Climate	Notation	Main Characteristics	Summer	Winter	Zonal Vegetation
Equatorial	I	Warm and wet year-round; no frost in lowlands	Inter-Tropical Convergence Zone (low pressure, esp. equinoxes)		Tropical rainforest
Tropical Summer-Rain	II	Distinctly wet summer and dry, cooler low-sun season	ITC	STHP	Raingreen forests/woodlands
- Savanna	II-III	Longer dry season, short wet	(less ITC, more STHP)		Savanna/thorn scrub
Subtropical Arid	III	Dry year-round, with cooler winter; various rain patterns	Subtropical High Pressure		Desert/semi-desert
Mediterranean	IV	Dry summer, winter rainfall (mild winter, spring growth)	STHP	Windward westerlies	Sclerophyll forest, maquis, garrigue
Marine West-Coast	Vm	Oceanic conditions year-round (wet, cool summer and winter)	Westerlies on windward coasts		Temperate rainforest
Warm-Temperate (east sides)	Ve	Warm/hot summer, mild winter, no true dry season	Convective rainfall	Frontal rainfall	Laurophyll/mixed forests
Typical Temperate (cf nemoral)	VI	Warm summer, frosty winter, no true dry season	Convective rainfall	Frontal rain/snow	Summergreen forest
Temperate Continental	VII	Warm summer with rainfall early; colder, dry winter	Convective rainfall	Frontal rain/snow	Prairie, steppe, dry grasslands
- Temperate Arid		[less, more erratic precipitation]	(same as VII)		Cold-winter deserts
Boreal	VIII	Cool/mild summer, severely cold winter; summer rain	Frontal rain + convection	Dry cold polar air	Boreal conifer forest, larch woods
Polar	IX	Cold summer, severe winter;	Dry cold polar air masses		Polar tundra

The climatic zones are those of Walter (1970), as expanded by Box (2002, in press; Box and Fujiwara 2013). The first five types (I through Vm) result from different seasonal patterns of control by the Intertropical Convergence Zone (ITC) or Subtropical High Pressure (STHP) belt, or both, which produce these climate types. The last two types result from high latitude and consequent low sun angle and low solar input, producing severely cold winters (except in oceanic areas) and short cool summers. The three climates in the middle (Ve, VI and VII) are differentiated less by global-scale circulation. All three receive mainly convective precipitation in the warmer seasons, often enhanced by local or regional monsoon-like effects, and are subject to frontal disturbances in the transitional and colder seasons. These climates differ mainly by latitude (relatively warmer or colder seasons) and by geographic position (relative position on a land mass), which results in more precipitation near oceans, especially near windward coasts, and much less toward continental interiors. The natural vegetation types corresponding to these climate types, called zonal vegetation, are given in the last column. Upland climatic zonation and vegetation belts can be understood as modifications of corresponding basic lowland zones.

Table 2: Major Upland Areas of the World

Mountains and Ranges	Geomorphic Form	Climate Zones
Appalachians	linear	VI
- Smoky Mtns	group	VI
- Adirondack, White mtns.	group	VI
- Gaspé	range	VI-VIII
Laurentian plateau	upland shield	VIII
Cascades	linear	Vm/VII
Sierra Nevada	linear	IV/VIIa
Rocky Mountains	cordilleras	VII, VIII
- Kaibab Plateau	plateau	VIIa
Sierra Madre Oriental/Occidental	linear	II
Andes	cordilleras	II-I-II-III-IV-Vm
- Altiplano	high plateau	II-III
Guyan plateau	upland	II
S Brasil plateau	upland	II-Ve
Scandes	linear	VI-VIII
Alps	arcs	VI
Pyrenes-Cantabrians	linear	VI
Spanish meseta	plateau	IV-VI
Apennines	linear	IV-VI
Central European uplands	variously oriented ranges	VI
Carpathians	arcs	VI
Dinarics + Pindus, Balkans	linear + spurs	VI
Urals	linear	VIII-VI
Atlas	ranges	IV
Ethiopian highlands	upland	II-Ia
E African uplands	upland	Ia
- Kilimanjaro, Mt Kenya	inselberg	Ia
Drakensberg	linear	II-Ve
Madagascar mountains	linear	II
Anatolian plateau	upland	IVc-VII
- Taurus	curvilinear ranges	IV-VII
Caucasus, Lesser Caucasus	ranges (short)	VI
Iranian Plateau	plateau	VII-III
- Elburz, Kopet Dagh	ranges	IVc
- Zagros	ranges	III-IVc
Hejaz	L-shaped	III
Central Asian ranges		
- Sayani Mtns	curvilinear range	VII-VI
- Altai	range	VII
Tien Shan system	range	VII
- Pamir	nodal mountains	VII
- Hindu Kush, Karakoram	ranges	VII
Tibet-China		
- Himalaya	linear	II-Ve
- Tibetan Plateau	plateau	Ve
- Kunlun, + Qilian-Altun	plateau ranges	Ve
- Yunnan plateau	plateau	Ve
- Emei-Shan	single	Ve-II
Qin Ling, + Daba, Taihang	ranges	VII
- Tianmu-shan, Huang-shan	single	VI-Ve
Nan Ling, + Wuyi	ranges	Ve

East Siberian ranges		
- Chukotkan, Kolymsk	ranges	VIII
- Verkhoyansk, Chersk	arc ranges	VIIIc
- Koryak-Kamchatka	linear	VIII
- Stanovoy-Dzhugdzhur	linear	VIII
- Yablonov	linear	VII-VIII
- Sikhote Alin	linear	VI
Khinggan (Greater, Lesser)	ranges	VI-VII
Changbai-Shan	range	VIc
- Taebaek (Korea)	range	VIc
Japan Alps	ranges	Ve-VI
- Fuji-san	inselberg	Ve
Deccan Plateau	plateau	II
- W and E Ghats	ranges	II
SE Asia		
- Annamiti (Vietnam-Laos)	cordillera	II
Bornean upland	branching ranges	I
- Mt Kinabalu	inselberg	I
Sumatra-Sunda	cordillera	I-II
Bismarck range	linear	I-II
Great Dividing Range	linear	II-Ve
Southern Alps (NZ)	linear	V-Vm

Cordilleras, arcs and most other linear systems are of tectonic origin, and many of these extend across multiple climatic zones. This is especially true of the Andes, Rocky Mountains and other cordilleras with mainly north-south orientation, as in the Western Hemisphere. Linear systems in the Eastern Hemisphere have resulted largely from collisions between Africa and Eurasia or from the collision of India with southern Asia. These mountains are generally oriented east-west and lie mostly within a single climatic zone, but some such systems lie far away from the areas of collision, as in eastern China and in western Europe. Broader upland areas and isolated single mountains lie mostly within a single climatic zone, including the Tibetan Plateau, remnants of old cratons such as the Laurentian and Guyana uplands, and inselbergs like Mt. Kilimanjaro.

Vegetation reflects climatic conditions in integrated forms, or pheno-physiognomies, which include aspects of both form (structure, finer morphology) and its seasonal variations. Vegetation also responds to critical threshold values and damaging temperatures, which preclude forms found elsewhere. The result is visible in the landscape as different (natural) vegetation types. Vegetation discontinuities, or at least transitions, thus represent useful landmarks for separating climate types in situations where climatic conditions are changing only gradually (i.e. most situations). The relationship between climate type and vegetation can thus be improved by tying some limits to known mechanisms of biotic limitation, as suggested in Table 3. Such mechanisms might include climatic criteria for biotic requirements like wood production or the production of viable seeds [e.g. 85, 32], and for damage to plants, as by intra-cellular ice formation [39] or damage to overwintering buds. This linkage provides a simple but biologically significant set of temperatures for delimiting the main climatic zones (Table 4). This does not change the basic concepts of climate types, as shown in Table 1, but it does provide a perhaps more consistent basis for separating climate types and for placing transitions geographically.

Table 3: Requirements and Limiting Mechanisms of Basic Plant Processes

Plant Process	Limiting Mechanisms and Critical Temperatures	References
Photosynthesis and growth	Temperatures above freezing or near freezing, at least at ground level (or in the soil) Cell and tissue hydration, and soil temperatures high enough (unfrozen, at least) for uptake of soil water	Larcher 1994, p 97 cf Cavender-Bares & Holbrook 2001
Wood Production	At least 1 month with effective mean $T > 10^{\circ}\text{C}$ (cf higher temperatures near ground level, exploited by low-growing plants)	Supan 1884, Andersson 1902 (see more references in Tuhkanen 1980)
Viable fruit/seeds	At least about 4 months with mean $T > 10^{\circ}\text{C}$ for typical temperate-zone trees Less for boreal broad-leaved trees, conifers and shrubs As little as 2 weeks for some herbaceous plants	Rubinstein 1924, Enquist 1924 (cf Walter 1968) Köppen 1884
Cold Damage to plant parts:		
Ice in wood cells	$T < -40^{\circ}\text{C}$ for trees and shrubs with ring-porous wood; diffuse-porous wood tolerates to about -70°C)	George et al. 1974, Arris & Eagleson 1989; cf Cavendar-Bares 2005
Ice in leaf cells	$T < -15^{\circ}\text{C}$ for evergreen broad-leaved trees and shrubs; mean $T < \text{about } -30^{\circ}\text{C}$ for conifer needles $T < -15^{\circ}\text{C}$ for evergreens, even in tropical mountains [deciduous plants of course are not affected]	Sakai 1971, Sakai & Weiser 1973; cf Larcher 2005 Cavender-Bares 2005 Sakai & Larcher 1987, Woodward 1987 García-Varela & Rada 2003
Damage to unprotected buds	$T < -3^{\circ}\text{C}$ for plants considered subtropical $T < 5^{\circ}\text{C}$ for plants considered tropical	Larcher 2005; 1994, p272; cf Sakai & Larcher 1987

Legend: T = short-term (e.g. overnight) air temperature, unless specifically denoted as mean temperature.

Table 4: Suggested Limits for Main Climatic Zones, based on Biotic Limiting Mechanisms

Polar zone	Fewer than 30 days $\geq 10^{\circ}\text{C}$, if any at all
Boreal zone	At least 30 days of mean $T \geq 10^{\circ}\text{C}$
Temperate zone	≥ 120 days (4 months) $\geq 10^{\circ}\text{C}$ mean
Warm-temperate zone	$T_{\min} \geq 0^{\circ}\text{C}$ and $T_{\text{abmin}} \geq -15^{\circ}\text{C}$ (and at least 4 months warm, i.e. mean $\geq 10^{\circ}\text{C}$)
Subtropical zone	$T_{\min} \geq -3^{\circ}\text{C}$
Tropical zone	No temperature ever below 0°C , perhaps not $< 5^{\circ}\text{C}$

These suggested limits follow directly from critical limiting temperatures for the production of wood, viable fruits and seeds, for damage to unprotected buds, and for lethal ice formation in plant tissues (see Table 3)

Abbreviations: T = mean temperature; $T_{\min}$ = mean temperature of the coldest month; T_{abmin} = absolute minimum temperature (lowest ever measured or expected)

What is "Lowland" and Where do Uplands Start?

That vegetation is different in uplands is evident [e.g. 19, 22, 62, 63, 71, 97-101, 104, 106]. The question of where uplands start is illustrated in Photo 1, which shows a stand of Scots pine (*Pinus sylvestris*) descending and becoming less dense toward the southwestern shoreline of Lake Baykal in south-central Siberia. A very similar scene could be found, for example, on the eastern edge of the Front Range of the Rocky Mountains in Colorado or in parts of the western

Himalaya (albeit with different pine species). As the ground turns upward, its steepness increases at first, due to accumulations of material at the base of the slope. Above that the slope becomes a more constant incline, decreasing again toward the top. The grassland area is quite small here at Lake Baykal and may be due to a local rain-shadow effect by the low range of hills to the west of the lake. Grassland covers a much larger adjacent area in Colorado. In both cases, though, the grassland changes upward into an open, and then denser pine stand. Where does the upland start – and why?

Initially one might seek both topographic and climatic criteria. Topographically, uplands appear to start where foothills start, or at “breaks” such as escarpments. Climatically, uplands may appear to start where gradients are steeper or begin to be steeper, or where temperatures or water balance reach critical values or thresholds. In the southern Appalachians, for example, evergreen *Rhododendron* appears abundantly in foothill forests (Photo 2) but not in the generally lower humidity of the adjacent piedmont. In particular, lowlands also lie below the level of steep gradients of improving water balance, with perhaps a disappearing dry season, and within atmospheric inversion layers (or below persistent cloud layers), as on islands and over cold currents. Lowlands may include piedmonts, even low foothills or tablelands, but not higher belts.

On the other hand, distinction of upland and lowland environments may be complicated in some areas by simultaneous, offsetting changes in elevation and latitude. For example, there is remarkably little change in the physiognomy of the tussock-steppe vegetation as one descends southward from the dry subalpine and alpine puna of southern Peru and Bolivia, with *Festuca orthophylla* and *Stipa ichu*, plus *Lepidophyllum quadrangulare* shrubs and small succulents like *Tephrocactus* [106 - pp. 250-256], to the much lower, very windy Patagonian steppe, with *Festuca pallenscens*, *Stipa speciosa*, and thorny *Mulinum spinosum* [104 - pp 711-718). Some other pairs of similar upland and lowland vegetation are suggested in Table 5.

Table 5: Corresponding Lowland and Upland Climatic Situations

Upland Situation	Corresponding Lowland Zone	Distinguishing Upland Factors
Alpine (temperate), also upland boreal	Polar zone	Daylength, greater surface warmth, length of T > 0°C period
Subalpine (temperate)	Boreal zone	Windier, may be cloudier, longer potential insolation
Montane (humid temperate) (e.g. mixed forest)	Sub-boreal transition	Greater average humidity?
Montane (humid subtropical) (laurophyll forest)	Warm-Temperate	Greater average humidity? Greater air mixing; less severe cold
High-montane/subalpine of tropical wet-dry regions	Humid subtropical/warm-temperate zones (laurophyll)	Water-balance seasonality
Upper-montane/subalpine of mediterranean climates	Lower uplands of temperate continental zones (e.g. dry pine)	Summer water balance
Dry tropical alpine/subalpine (e.g. Peruvian dry puna)	Maritime dry temperate (e.g. Patagonia)	Daylength and sun angle [seasonality?]

Corresponding uplands differ almost always from otherwise similar lowlands by having more inclined, usually coarser-textured and patchier substrates, which are thus generally better drained, drier, and more heterogeneous (but including wetter areas where water runs in). Climatically, corresponding upland climates generally occur at lower latitudes and thus have less extreme daylength regimes (especially in high latitudes) and more intense insolation (especially in lower latitudes). The latter, in particular, accentuates the heterogeneity of surface environments. Other climatic differences are less consistent or clear, but uplands may be generally windier and consequently have less extreme low temperatures, due to atmospheric mixing; uplands thus do not have continuous permafrost over large areas, and they do not have large expanses of saturated soil. Uplands also have vertical lapse-rate regimes, as well as

vertical precipitation gradients, and neither of these vertical patterns is necessarily monotonic. Finally, uplands and corresponding lowland environments sometimes merge (e.g. humid temperate montane and sub-boreal mixed forest in the northern Appalachians or northern Europe).

Relative to lowlands, upland environments (see, for example, Crawford 2008, chapter 10) are usually:

- cooler (except in inversion layers),
- windier,
- cloudier and more humid,
- wetter, i.e. with more precipitation (at least up to the cloud layer),
- sunnier (above the cloud layer), with more intense solar radiation,
- faster to heat up and cool off, especially near the ground surface,
- rougher (rockier) and more inclined, both of which features generally increase upward,
- more rapidly drained, with shallower, patchier soils, and
- subject to more equitable daylength regimes than their high-latitude lowland analogs.



Photo 1: Lower Pinewood Boundary on the West Side of Lake Baykal (southern Siberia)

The pine is *Pinus sylvestris*, which forms dense to more open stands across much in Siberia. Here, near Listvyanka (southwestern shore of Lake Baykal), the pine becomes gradually less dense and eventually disappears completely in the drier lee of the low hills on the west side of the lake. A similar natural ecotone can be seen in foothills of the Rocky Mountains in Colorado.

Uplands are cooler because Earth's atmosphere is heated from the bottom up, by solar energy absorbed at the (lowland) surface and transmitted upward by re-radiation, convection, conduction (more slowly), and release of latent heat when water vapor in the rising, cooling air condenses. On the other hand, near the ground surface, uplands may heat up faster than lowlands because of their more intense solar irradiation; uplands cool off faster at night (radiative cooling), because there is less atmosphere above to absorb outgoing longwave radiation.



Photo 2: Riparian *Rhododendron* in foothills of the southern Appalachian Mountains

Stands of several evergreen *Rhododendron* species, as well as *Kalmia latifolia*, are common in foothill forests of the southern and central Appalachians (North Carolina and Virginia), and not only along streams. These species usually do not occur in the deciduous forest of the adjacent piedmont, perhaps due to its generally lower humidity.

Uplands are usually windier because of less friction with Earth's surface. Uplands are more humid because of the adiabatic cooling of rising air, which reduces its saturation capacity (raising relative humidity), and cloudier because rising air eventually cools to its dew-point temperature and water condenses out as clouds. Wind blowing against mountain ranges may enhance this orographic effect of upward water transport and cloud formation, resulting in greater precipitation as the accumulating water droplets become too heavy to remain aloft. But precipitation increases only up to cloud level, above which it may be drier again.

In terms of topography, uplands (especially higher mountains) normally have rougher (rockier) and more inclined ground surfaces than do lowlands. Mountain soils are usually shallower and coarser, due to the removal of finer-grained components by erosion, and are usually very patchy, with large differences in depth, texture and dryness over quite small distances, due to the overall roughness of the substrates.

Finally, compared with what may appear to be analogous climatic or vegetational situations at lower elevations (but inevitably higher latitudes), daylength regimes in mountains are usually significantly more equitable.

Upward Climatic Patterns

Uplands are recognizable because the natural landscape changes significantly, as from evergreen broad-leaved forest to semi-evergreen or deciduous forest, or from deciduous to mixed

forest. In tropical wet-dry regions, uplands can be seen as beginning where the dry season is ameliorated enough for raingreen deciduous woods to become semi-evergreen.

Topographic criteria may work in the field or in more sophisticated GIS operations, from broad down to landscape scales. For generalizing and modeling, though, climatic criteria are probably needed to delimit the start of the uplands.

Temperature:	critical levels may apply only to alpine belts and some subalpine (vegetation of temperate affinities)
Humidity:	probably for most climates – but data are rarely available
Water balance:	can perhaps be estimated, but what aspects?

Climatic criteria may be less applicable at landscape scales, but even a modest rise in elevation can increase humidity and reduce the evaporative water loss by plants – or at least shorten the duration of a water-stress period.

Rising air cools automatically by about 9.8°C per 1000m upward (about 1°C per 100m), since lower surrounding pressure permits the air mass to expand and spread its internal thermal energy over that larger air volume. This 'dry adiabatic lapse rate' is constant until dew-point temperature is reached, beyond condensation in the cooling air releases latent heat, which partly offsets the adiabatic cooling. The resulting 'wet adiabatic lapse rate' is less than the dry rate but is no longer a constant, depending now on the actual temperatures involved.

Warmer air can "hold" (equilibrium mixing ratio) more water vapor than cooler air, and this saturation vapor capacity roughly doubles for each 10°C increase in temperature. This means, conversely, that air 10°C cooler can "hold" only half as much water vapor and its relative humidity is doubled, provided that there is no significant change in the actual amount of water vapor in the air (and that saturation has not been reached). As a result, the saturation deficit of rising air (saturation minus actual water-vapor content) is decreased significantly within a few hundred meters upward, reducing potential evapotranspiration in almost linear proportion. This can reduce the evaporative water stress on plants significantly – or at least shorten the duration of stress.

Humidity begins to increase and saturation deficit to decrease as soon as air is pushed upward. The effect after only 100m can be a reduction of 5-10% in saturation deficit, which may already be significant for some plants occurring near their climatic limits. Of course relative humidity and saturation deficit also vary quite significantly over the course of 24 hours, following temperature. If afternoon is only 10°C warmer than dawn (when temperature is normally lowest), and if dew had formed overnight (i.e. 100% relative humidity was reached, at least at ground level), then the afternoon relative humidity would fall to about 50% and the saturation deficit (i.e. water stress) would become significant. It is not difficult to make these calculations and to apply them in models. The problem is obtaining the humidity data with which to begin a calculation.

Apart from rising air, the air at different levels above the lowland surface is also cooler upward because the atmosphere is heated from the bottom up (*vide supra*). In the free atmosphere, this rate of temperate decrease upward, the 'environmental lapse rate' (ELR), is about 6.4°C per 1000m upward as a world average. Along slopes the cooling rate is less, though, due to the effect of the heated slopes. This has been called the 'terrestrial lapse rate' (TLR, Meyer 1992) and is around 5.6°C per 1000m on average, but varies more than does the ELR.

The effect of elevation and these cooling processes to reduce dryness in the montane belt is illustrated by a simple calculation for Jianfengling [ling = mountain range], a mountain in central Hainan island, near Dongfang (18° N), where there is a distinct dry season (Box et al. 1989). For lowland Dongfang, the mean monthly temperature (T, °C), average precipitation (P, in mm), potential evapotranspiration (PET, estimated by the method of Holdridge 1959) and P/PET ratio (MI, moisture index) are as follow:

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
T:	19.4	20.5	22.4	26.0	28.6	29.4	29.4	28.9	28.2	26.4	23.7	19.9
P:	8	15	20	31	83	134	117	231	182	107	21	9
PET:	95	101	110	128	141	144	144	142	139	130	116	98
MI:	.08	.15	.18	.24	.59	.93	.81	1.63	1.31	.82	.18	.09

Mean annual temperature is 25.2° C, average annual precipitation is 957mm, and estimated annual PET is 1488mm, for an annual P/PET ratio of 0.643. Dongfang, and much of interior and western Hainan, thus have a fairly typical tropical summer-rain climate, with mainly dry raingreen woods and some areas of savanna as the natural vegetation. The dry season ($P < PET$) extends from October through July (10 months), with a short wet season in August-September, estimated as 2.4 months in length. The P/PET ratio in the driest months (December and January) is around 0.09, and 6 months have 31mm or less precipitation, 5 months have 21mm or less.

At 890 m, however, we found a semi-evergreen montane forest dominated by *Dacrydium pierrei* (Podocarpaceae), along with *Xanthophyllum hainanensis* (Xanthophyllaceae), laurophyll trees *Lithocarpus fenzelianus* (Fagaceae) and *Polyspora* (=Gordonia) *balanse* (Theaceae), and various other laurophyll trees in the canopy and understorey. In the absence of climate data at 890m, the corresponding monthly climatic values were estimated, based on the world-average ELR [15]. These values are re-estimated here (via program POLATE, Box, unpublished), based on the more applicable TLR, i.e. an upward temperature decrease of 5.6°C per 1000m and a perhaps better estimate of the cloud-base height, which varies monthly, based on estimated mean relative humidity in the lowland:

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
T:	14.5	15.6	17.5	21.1	23.7	24.5	24.5	24.0	23.3	21.5	18.8	15.0
P:	11	21	28	44	117	189	165	325	256	151	30	13
PET:	71	76	86	103	116	120	120	118	114	105	92	74
MI:	.16	.28	.33	.42	1.01	1.57	1.37	2.76	2.24	1.43	.32	.17

At 890m, mean annual temperature dropped to an estimated 20.3° C, average annual precipitation had risen to about 1350mm, and estimated annual PET had dropped to about 1200mm for an annual P/PET ratio of 1.13. Although precipitation, both estimated and in reality, is not much greater in the driest period, the lower temperatures mean that the number of "wet" calendar months ($P > PET$) rises from 2 at Dongfang to 6 (May through October) at 890m, for a potential growing season (warm plus $P > PET$) of almost 6 months, more than double the length of that in the lowland. Perhaps more importantly, the length of the dry season had dropped from 10 to 6 months, and only the two driest months show a severe moisture deficit ($MI < 0.2$). The weak link in both the 1989 and current calculation is, of course, the estimated rate at which precipitation increases upward. Even so, both calculations suggest that the dry season is ameliorated significantly at less than 1000m above the lowland. The overwhelming evergreenness of the montane forest vegetation suggests that most trees are able to find enough soil water to meet evaporative losses through the shorter dry season, which occurs in the cooler part of the year.

The effect of elevation on vegetation is suggested more generally in Table 6, for representative sites in those main climate types that have a significant dry season (II, IV and VII) or are dry all year (III and VIIa). Effects are shown for upward shifts of 1000m and 2000m, derived from corresponding lowland values and based on an average terrestrial lapse rate of 5.6°C per 1000m. Dry periods are shown as the number of consecutive (calendar) months with average P less than PET. Wet periods represent the longest continuous [above-freezing] period with P > PET (calculated digitally, not as calendar months). The potential growing season is the intersection of the wet period and the period with mean temperature above an appropriate growing-season threshold, with one month added (unless it is too cold) for growth using stored soil moisture.

All sites show at least one fewer dry months at 1000m, and most show a further decrease at 2000m. For the tropical wet-dry sites, the wet period and potential growing season increase significantly at 1000m and even more at 2000m, suggesting that raingreen (deciduous) vegetation could be replaced by semi-evergreen vegetation already at 1000m (as at Jianfeng) and by largely evergreen vegetation at 2000m, at least for the wetter site (Banyo). A much more modest potential growing season appears upward for the hot-desert site (Menaka), and this seems similar to situations described for other, isolated upland areas in the Sahara, such as the Hoggar (Ahaggar) and Tibesti massifs [81, 90, 53; cf 106 - pp 637-642, 38]. Even at 2000m, temperatures at these sites remain well above limiting levels.

Cold does play a role in uplands at the mediterranean and dry-temperate sites. Although the number of dry months (above freezing) decreases upward, the length of the wet period may not increase because only sufficiently warm periods are counted. Oddly, in some situations such as mediterranean mountains, the length of the potential growing season may increase slightly upward but then become suddenly shorter when freezing or other low temperatures appear, usually at some lower to middle elevation; the growing season may increase gradually again above this level, as drought is ameliorated further. This "double bind" situation in mediterranean climates was illustrated graphically by Miller and Hayek (1981, p. 40). In Table 6, only Santiago is not affected by this, since its temperatures are still above freezing at 1000m (but not at 2000m). A similar gradual reduction of potential growing season is shown for Tashkent and Lahontan. Even so, the degree of drought at these sites (as suggested by dry-period P/PET) is less, and some plants may continue to grow even with the slightly shorter periods of good growing conditions.

From these considerations it does appear that water balance may be the main limitation mechanism for upward vegetation zonation in mountains, including the differentiation of the first upland situations from the adjacent lowlands. Limitation may, however, also involve more subtle aspects of water balance, such as minimum soil moisture during the driest period; soil moisture during spring or other critical period (such as reproduction); and periods of rapid soil drying. The critical limiting factors may be different in different climate and vegetation types. Nevertheless, the point of fastest change in one or more critical aspects of water balance may represent a good estimate of the start of the "uplands".

Higher Vertical Zonation

From the beginning, the concept of altitudinal belts in mountains was more a vegetation concept than a climatic concept. Belts were recognized by their vegetation [78]. Inclusion of

upland zonation in a global system first requires identification of the lowland climate zone or zones within which the particular upland rises [cf 107]. Walter used X to denote mountain climates in general, so mountain climates can be denoted by adding the corresponding lowland zone, e.g. X(VII) for uplands in a Temperate Continental climate. Further upland zonation, though, requires delimitation by differences in vegetation. The region above treeline, up to the line of more or less permanent snow (if the mountains are high enough), is normally called the alpine belt and has treeless vegetation somewhat analogous to polar landscapes. The first belt below treeline (or dry-climate equivalent) is the subalpine belt, and the first quite different landscape up from the lowland, such as woodland occurring above grassland, is called the montane belt. For example (see Photo 3), in the central Rocky Mountains of Colorado, in a temperate continental (VII) climate with short-grass steppe in the lowlands:

- the montane belt is dry pine forest (mainly *Pinus ponderosa*);
- the subalpine belt is a less-dry, mainly non-pine conifer forest (mainly *Picea* and *Pseudotsuga*); and
- the alpine belt is dry alpine "tundra" dominated at maturity by *Kobresia*.

In middle and higher latitudes, conditions and vegetation in subalpine belts may be somewhat similar to the lowland boreal (VIII) zone; humid montane belts may have mixed or deciduous forests similar to those in the nemoral-boreal (VI-VIII) transition zone.

In equatorial climates (see Photo 4), the zonal vegetation in the lowlands is tropical rainforest. Above that is 'montane rainforest', in which trees are still quite tall, precipitation may be even greater than in lowlands, but cooler temperatures upward may reduce the number of epiphytes and lianas. This pattern was described quite early [e.g. 9, 10; cf 82, 112]. A more or less permanent cloud belt occurs around 1000m above the base of the mountains, within which the 'cloud forest' is stunted by high humidity and lack of sunlight, and is covered in mosses [41]. This is the 'elfin woodland' of Beard (1944) and the *Elfenwald* of earlier German descriptions. Above the cloud level there is usually a subalpine belt that is not as wet and which represents the transition to alpine. In the alpine belt conditions are harsh, with temperatures above freezing every day and below freezing essentially every night [cf 86]. The result can be seen in the Andes [e.g. 26] and East Africa [e.g. 42], perhaps also New Guinea, as an oft-cited example of convergent evolution, namely the caulirosette forms with terminal tufts of large, quite fuzzy (white-pubescent) but otherwise soft leaves, that open during the day and close at night, thus protecting the growth points against the freezing cold [e.g. 106; cf 36].

In spite of the large degree of geographic and ecological complexity, there does seem to be a global pattern in the potential vegetation of the different altitudinal belts in the different climate zones, as suggested in Table 7. Climates that have forest in the lowlands may give way to more (thermally) seasonal forests in mid-latitude mountains or to less seasonal forests in the tropics, where the dry season may be ameliorated upward. Drier climates generally show denser vegetation upward, with grassland or dry scrub replaced gradually by woodlands and perhaps forest, before tree growth is precluded by the low temperatures of the alpine belt. As in the lowlands, the climates with both wet season and dry season are especially interesting, since the dry season may be ameliorated quite significantly upward – or may be truncated even below the alpine belt by shorter warm seasons. Alpine belts are all treeless but can have quite different vegetation in the different climate zones, from the caulirosette *páramo* scrub of equatorial climates to the wetter and drier *puna* grasslands (II and III), the thorn scrub of Mediterranean alpine belts, and the more tundra-like vegetation in the middle and boreal latitudes.

Table 6: Effect of Upward Shift on Dry Season in Different Climates

	Tropical Wet-Dry (II)		Subtropical	Mediterranean (IV)		Temperate	Temperate
	wetter	drier	Arid (III)	wetter	drier	Dry (VII)	Arid (VIIa)
	Banyo	Bangalore	Menaka	Siena	Santiago	Tashkent	Lahontan
	(1098m)	(920m)	(278m)	(348m)	(550m)	(428m)	(1267m)
Up 2000m (below P peak)							
- Tmin	9.8	9.5	12.4	-6.6 (5)	-3.4 (4)	-11.5 (5)	-11.0 (6)
- dry-period P/PET	0.63	0.36	0.20	0.81	0.17	0.21	0.23
- dry months	2	4	10	2	6	4	5
- wet period	10.0	7.5	1.6	3.8	2.0	2.0	1.7
- growing season	10.0 +	7.4 + 1	1.6 + 1	0.4 + 1	--	--	--
Up 1000m (lapse = 5.6°C per 1000m):							
- Tmin	15.4	15.1	18.0	-1.0 (1)	2.2	-5.9 (3)	-5.4 (3)
- dry-period P/PET	0.44	0.31	0.17	0.59	0.15	0.18	0.18
- dry months	3	5	11	2	7	6	7
- wet period	8.7	6.7	0.5	5.2	5.0	2.4	1.8
- growing season	8.7 +	6.7 + 1	0.5 + 1	1.3 + 1	1.9 + 1	0.3 + 1	--
Lowland							
- Tmin	21.0	20.7	23.6	4.6	7.8	-0.3 (1)	0.2
- dry-period P/PET	0.39	0.41	0.11	0.56	0.16	0.15	0.13
- dry months	5	9	12	5	8	7	10
- wet period	6.9	3.2	--	7.2	3.3	2.8	1.9
- growing season	6.9 + 1	3.2 + 1	--	1.4 + 1	0.3 + 1	0.5 + 1	--

Effects are shown for upward shifts of 1000m and 2000m, at sites representing wetter and drier variants of climates that have significant dry seasons, plus subtropical arid (dry all year). Upshift effects are derived from the lowland values, based on a terrestrial lapse rate (environmental lapse rate along a land surface, cf Meyer 1992) of 5.6°C per 1000m; this rate may be less, especially in more oceanic areas (cf Crawford 2008).

Tmin is the mean temperature of the coldest month (numbers of below-freezing months in parentheses). The dry period is defined by $P < PET$ (P = average precipitation, PET = estimated potential evapotranspiration) and described by P/PET during that period and the number of consecutive dry (calendar) months represented. The (main) dry period may come in summer (IV, VII, VIIa), winter (II) or all year (III), but dry-period temperatures are always well above limiting (even above 9°C at +2000m). Some climates, such as IV and VIIa, may have a main growing season in spring and another period of activity in autumn, before it becomes too cold. The wet period is the duration of the longest continuous period with $P > PET$, and growing season is the intersection of the wet period and the period with mean temperature above the growing-season threshold (estimated from overall annual climatic conditions and decreasing with increasing altitude). One month is added (unless temperatures are too low) for the use of soil moisture to support plant activity. Period lengths were determined from the mean monthly data by PERIOD (Box, unpublished).

Sites were chosen to be representative of their climate types but also close to highland areas with which the upshift results can be compared. The wetter tropical wet-dry (II) site is Banyo, in the lower mountains of Cameroon; the drier II site is Bangalore, a dry-deciduous forest site on the Deccan Plateau of south-central India. The subtropical arid (III) site is Menaka, in the Sahara desert of Mali. The Mediterranean (IV) sites are Siena in Tuscany and Santiago in central Chile. The temperate continental (VII) site is the dry grassland area of Tashkent in Middle Asia, and the temperate arid (VIIa) site is Lahontan, at the east base of the Sierra Nevada in the Great Basin desert of Nevada (western USA). Roman numerals denote the climate types of Walter (1970, 1977).



Photo 3: Altitudinal Vegetation Belts in the Equatorial Climate of the Venezuelan Andes

Upper left: Cloud belt above Mérida, situated at about 1700m elevation. Upper right: Montane rainforest at lower limit of the cloud belt (about 2500m). Lower left: Upper limit of the cloud belt, with subalpine scrub including plants from the alpine páramo. Lower right: *Espeletia* scrub of the alpine páramo, also with small-leaved evergreen shrubs, grasses and forbs. The caulirosette growth form of *Espeletia*, which opens its leaves in the daytime and closes them at night, is considered to be an adaptation to the equatorial alpine climate, which has no dry season but temperatures that are above freezing every day and below freezing almost every night. Similar plant forms occur in the equatorial mountains of East Africa

There are complications to this systematic zonation, however, and an indispensable starting point for considering these problems is the very useful review by Sandro Pignatti (1980) on the problem of altitudinal belts and vegetation. The basic terminology was established already in 1910, but always with vegetation taken as the starting point, since it is recognizable in the field. A vegetation zone was defined as an areal unit recognizable through a broad-scale comparison of lowland landscapes, characterized by a uniform vegetation region and a somewhat homogeneous bioclimate (paraphrased from the German definition given by Pignatti 1980. An

altitudinal belt was defined as an areal unit in mountains, recognizable by a uniform vegetation belt and with a somewhat homogeneous bioclimate. It was also recognized also that:

- similar vegetation can be characteristic of altitudinal belts or lowland zones;
- the lowest vegetation belt is equivalent to the surrounding [lowland] vegetation zone; and
- vegetation zones are based on sea-level but can be extended to mountains in warmer regions.

The problem comes when one observes that the concept of altitudinal belts may not always be consistent. Even within Europe, the vegetation just below treeline, i.e. in the subalpine belt, is not always a conifer forest but may sometimes be beech forest (*Fagus sylvatica*), as in parts of the Apennines (Italy) and the Dinaric Alps and other Balkan mountains [78]. One could say that these exceptional areas are not subalpine belts in the normal nemoral (VI) climate but rather in a more submediterranean or even mediterranean climate. Even so, the occurrence of beech versus conifers seems also to be related to calcareous versus acidic substrates.

That vegetation is the criterion for recognizing altitudinal belts was stated flatly by Braun-Blanquet: "altitudinal belts are bioclimatic vegetation belts" [78]. Every altitudinal belt corresponds to a vegetation belt, but, as suggested above, not every vegetation belt corresponds necessarily to an altitudinal climatic belt. This view can be summarized in the following simple table that compares levels of abstraction in the different concepts:

Phenomenon	Existence	Recognition
vegetation areas/ formations	concrete	directly recognizable
altitudinal belts and vegetation zones	concrete (areas)	recognized only through their vegetation
bioclimates	abstract (not concrete)	not recognizable directly, only through measurements

Even so, if we are to have a global classification system, it probably needs some method of climatic quantification separate from the recognition of vegetation types.

Bioclimatic belts normally occur at higher elevation in more continental areas (*Massenerhebung* effect), since belts (outside the tropics) are determined largely by summer temperatures [cf 44, 48, 110, 73, 69, 56, 49]. This was documented in East Asia, for 20 mountain situations all along 30°N latitude, by Fang et al. (1996), who found the upper limits of the main vegetation belts to be about 1500m higher in western China than in the east (which included some coastal and island situations). The upper limits of the various belts were also related to thermal parameters, in particular to two temperature-sum metrics for which altitudinal boundaries had already been postulated by their authors. These temperature metrics were the biotemperature of Holdridge (1947), i.e. the sum of mean monthly above-freezing temperatures, divided by 12; and the Warmth Index of Kira [50-52], i.e. the sum of mean monthly temperatures above 5°C. The average values found by Fang et al. [34] for the upper limits of the various belts were as follow:

	Biotemperature	Warmth Index
Alpine	2.3°	4.6°
Subalpine	3.7°	14.1°
Cool-temperate	8.0°	53.3°
Warm-temperate/subtropical	11.1°	81.8°
Tropical	19.2°	170.6°

Cool-temperate, in this sense, is normally associated with temperate deciduous forests and warm-temperate/subtropical with evergreen broad-leaved ("laurel") forests; these can perhaps be interpreted as upper and lower montane belts, respectively. Fang et al. found further that the limit of evergreen broad-leaved trees appeared to occur at 9.6°C for the warm-month mean, 0.9°C for the cold-month mean, -8.8°C for Kira's Coldness Index, and -14.1°C for absolute minimum temperature. These values are similar to values found elsewhere, at least in the Northern temperate zone.



Photo 4: Altitudinal Vegetation Belts (central Rocky Mountains) in the Temperate Continental Climate of Colorado and Adjacent Areas

Upper left: Great Plains grassland, shown here in southwestern South Dakota but occurring westward right up to the eastern edge of the Rocky Mountains. Upper right: the montane belt with *Pinus ponderosa*, the montane-belt dominant in Colorado and over much of the interior western USA (shown here near the North Rim of the Grand Canyon in northwestern Arizona, at about 2100m). Lower left: the subalpine belt, with more moist-demanding conifers such as *Picea engelmannii* and *Pseudotsuga menziesii* (shown here in the main chain of the Rocky Mountains in Colorado, at about 2500m above Boulder). Lower right: the alpine belt of "dry alpine tundra" dominated at maturity by *Kobresia* but with many colorful forbs in earlier stages (shown here higher above Boulder, at a bit over 3000m; photo in latter July, at time of peak blooming). This last panel also illustrates a problem in vertical zonation, namely the widespread occurrence of treeless flat valleys surrounded (at the same or somewhat higher elevation) by remnants of the subalpine conifer woods. This phenomenon is attributed to cold-air drainage and consequently shorter effective growing seasons in the valleys.

Table 7: Zonal Vegetation of Altitudinal Belts in the Different Climate Zones

Zone	Lowland	Montane Belt	Subalpine Belt	Alpine Belt
Equatorial	Tropical rainforest	Montane rainforest: tall but fewer epiphytes	Cloud forest: short, stunted, with many epiphytes	Páramo scrub, with tuft treelets
Tropical Wet-Dry	Raingreen forests/dry woods	Raingreen to semi-evergreen forest	Low, often evergreen forest	Moist puna grassland, large tussocks
	Savanna/thorn scrub	Montane to subalpine grassland		
Subtropical Arid	Desert/semi-desert	Semi-desert scrub, with more succulents	Semi-desert scrub, more succulents	Dry bunch-grass puna
Mediterr.	Sclerophyll woods	Dry conifer woods, especially pines	More mesic woods/forest especially conifers	Thorn scrub, often with cushion forms
	Tall sclerophyll forest	Giant monopodial redwoods/eucalypts		
Marine W-Coast	Temperate rainforest	Montane rainforest (mixed in N Hem)	Wet alpine forest (conifer in N Hem)	Wet alpine tundra
Warm-Temperate	Evergreen broad-leaved forest	Laurel forest (often smaller leaves)	Evergreen mixed forest (broad-leaved + conifer)	Moist alpine tundra
Typical Temperate	Summergreen forest	Mixed forest: broadleaf decid + temp conifers	Conifer forest, with boreal or boreal-like species	Alpine tundra
Temperate Continental	Grassland	Dry conifer forests/woods, esp. pines in N Hem	Mesic conifer forests, with boreal-like species	Dry, grassy alpine tundra
Boreal	Conifer forest	[already subalpine]	[already alpine]	Polar-like tundra but with more dwarf shrubs
Polar	Tundra	[already subnival]	[already nival]	Moss-lichen cold-desert

Base climate zones and zonal vegetation are given in Table 1, following the expanded Walter climate types.

The most obvious altitudinal vegetation boundary is usually the treeline, which separates the subalpine and alpine belts. As for polar treeline, the alpine treeline has long been associated with the 10°C isotherm for mean temperature of the warmest month [18, 55, 27, 47, 40, 102, 76, 34]. A short comparison by Körner (1998), however, showed treeline occurring at warm-month means ranging from 5.6°C on Mt. Wilhelm (New Guinea) to 11.2°C in Siberia [70] and 11.4°C in the Cairngorms [40], and as high as 12.4°C in Hokkaido [76] and 13.0°C in the Hida Mountains (southern Japan), [93]. A more recent worldwide comparison by Körner and Paulsen (2004) found treeline to be correlated better with values of mean growing-season temperature, varying from 5-6°C in equatorial regions to 7-8°C in temperate and mediterranean regions. A more critical threshold factor, however, especially in the tropics, may still be mean soil temperature, perhaps a 0°C isotherm (which varies less over a year than in higher latitudes).

Further down, the montane and subalpine belts seem to be separated not by temperature but by water-balance factors, such as the degree of soil dryness in summer or the occurrence and relative elevation (above mountain base) of a cloud belt. Some of the climatic criteria needed for delimitation of the different altitudinal belts, in a general way, are suggested in Table 8. The alpine belt is delimited by various aspects of temperature, especially as they may affect the length of the

potential growing season and the occurrence of damaging cold during the growing season [cf 96]. The subalpine belt is delimited (from below) by the disappearance of some critical degree of water stress – a criterion that may need to be handled differently in tropical mountains, as suggested above by the persistent cloud belt (e.g. Photo 4). The montane belt is differentiated (from below) by water-balance criteria in dry climates and by temperature in humid climates. An additional, distinctly different upper-montane belt may occur in some mountains and cannot be generalized here. To the extent that a recognizable foothill (colline) belt occurs, it is probably to be differentiated (from the adjacent lowland) by humidity – except in cold valleys or basins with quite significant and persistent cold-air drainage.

Table 8: Criteria for Delimitation of Altitudinal Belts in Mountains

Belt	Recognition	Climatic Correlates
Alpine	Above treeline, up to permanent snow	Warmest-month mean temperature around 10°C (lower in oceanic areas in tropics, mean [annual] soil temperature around 0°C);
Subalpine	First belt below treeline (or dry-climate equivalent)	Disappearance of water stress, except short periods in summer, e.g. spruce-fir forest above pine
[Upper- Montane]	If another landscape appears before subalpine	[idiosyncratic]
Montane	First different landscape up from the lowland	Dry: Ameliorated water balance, e.g. woodland above grassland Humid: Colder winters, e.g. mixed forest above deciduous
Colline	Foothills, with gradually increasing humidity and vegetation density	Dry: Increasingly dense matorral, or maquis above garrigue Humid: Appearance of laurophylls or "northern" taxa in deciduous forest

Altitudinal belts are recognized by their vegetation but may have consistent climatic correlates or determinants. The vegetation of each belt is significantly different (except perhaps in extremely arid mountains with essentially no vegetation). For example (see Photo 4), in the central Rocky Mountains in the Temperate Continental (grassland) climate of Colorado:

- the montane belt (above short-grass steppe) is dry pine forest, of mainly *Pinus ponderosa*;
- the subalpine belt is a less-dry, mainly non-pine conifer forest, of mainly *Picea* and *Pseudotsuga*; and
- the alpine belt is dry alpine "tundra" dominated at maturity by *Kobresia*.

Three main belts are usually recognized (alpine, subalpine and montane), with upper-montane and foothill (colline) belts added as useful.

Variation in Temperature Lapse Rates

The world-average value of the environmental lapse rate in the free atmosphere is about 6.4°C per 1000m. The lower values found along mountain slopes, due to surface heating, are called the 'terrestrial lapse rate' (TLR), and a world average has been estimated as about 5.5°C per 1000m [71, 72, 113; cf 23]. This TLR value varies, though, in different climates. It is generally highest in subtropical deserts in summer (due to intense surface heating) and lowest in boreal-polar areas in winter [7].

Temperature lapse rates may also vary significantly within a climatic region. In principle, lapse rates should be greater ("steeper") with strong surface heating but also in oceanic and other cloudy, higher-latitude areas. Lapse rates are known to be steep in northwestern Europe, due to advection of cold polar air aloft [7; cf 23], and may be inverted in boreal-polar areas and in

mountain valleys. Fang and Yoda (1988) calculated the 'altitudinal lapse rate of temperature' in China based on data (1951-80) from 671 meteorological stations. They found the lapse rate to decrease inland from the coast, reach a minimum about 600 km inland, and then to increase again, to a maximum at about 1600-2000 km inland. This pattern of higher lapse rates in coastal areas, as in northwestern Europe, and may occur even in the tropics. For example, the Wikipedia page for 'Massenerhebung effect' (retrieved 30 Sept. 2014) refers to "moss forest" on Borneo occurring at 900m at Gunung Palung near the coast, at 1200m at Gunung Mulu inland, and at 1800m on Mt. Kinabalu.

In addition to a general increase at higher elevations inland [cf 33], lapse rates may also vary with elevation on individual mountains. Alisov et al. (1956) documented a more or less normal lapse rate above 1000m at Alma Ata (Uzbekistan, now Almaty) but a moderate to strong temperature inversion below 1000m, especially in winter. A general pattern of smaller lapse rates on lower mountain slopes was suggested by Yoshino (1975). Variation of TLR values with topography was studied in detail, in the western USA, by Wolfe (1992).

Finally, temperature lapse rates are usually higher in summer than in winter. On the Zugspitze (northern Alps), Hendl (1966) estimated the lapse rate as 3.9°C per km in winter and 6.3°C per km in summer, with an annual average of 5.2°C per km. For the Hoggar massif in the central Sahara, Barry (1981) estimated lapse rates of 5.0°C per km in winter and 8.0°C per km in summer. For Almaty, Alisov et al. (1956) estimated a seasonal variation from 4-6°C/km in winter to 5-8°C/km in summer (above 1000m). In their study of lapse rates in China, Fang and Yoda (1988) found lapse rates to be not only smaller in winter but also more variable, and greater but less variable in summer. They estimated lapse rates of 2.4-5.3°C per km in winter and 4.6-6.5°C per km in summer, with an annual average of 3.2-5.9°C per km (decreasing inland and then increasing again). These "local and seasonal variations of temperature lapse rate" were also well recognized by Fang et al., 1996.

Probably no group has estimated lapse rates in mountains more than the Chinese, whose literature provides values not only by months of the year but also by elevation and aspect on individual mountains [e.g. 95]. The problem of lapse-rate variation is illustrated by data in Table 9, from Leigong-Shan, a mountain in Guizhou, in the humid-subtropical climate of southern China (Leigongshan Committee 1989). Note that the lapse rates are highest in summer, generally lower in winter, and are higher at mid-elevation and less steep both above and below.

Temperature lapse rates may also vary on different sides of a single mountain. The effect of aspect on temperature patterns is suggested in Table 10, which also shows more detail on the seasonal variation (also from Leigong-Shan, see Leigongshan Committee 1989). Both the highest and lowest lapse rates are shown to occur on the north side of the mountain. As in Table 9, the monthly data in Table 10 also show highest lapse rates in summer and low rates in winter.

Finally, variation with elevation of some temperature extremes is shown in Table 11 for Lei-Shan, another mountain in Guizhou (Leigongshan Committee 1989). The data here do not extend as high upward but do suggest a pattern. Mean annual temperature decreases upward, as expected, but mean minimum temperature (T_{min}) decreases less and is actually slightly lower at 286m than at 500m. Absolute minimum temperature (T_{abmin} , the lowest ever measured) shows a different pattern, lowest at 641m and less cold at 840m. This could be due to cold-air

drainage or to greater mixing at higher elevation. Note, however, that the record is short, only about 20 years.

Table 9: Seasonal Lapse Rates of Temperature on Mt. Leigong-Shan (Guizhou)

	January	April	July	October	Year
1600-2179m	0.36	0.38	0.50	0.40	0.43
1100-1600m	0.54	0.54	0.50	0.50	0.52
640-1100m	0.39	0.39	0.52	0.39	0.43

Temperature lapse rate is the difference per vertical distance upward and shows a world average value in the free atmosphere of 6.4°C per 1000m. Due to surface heating, though, the Terrestrial Lapse Rate (TLR) along slopes is normally less, with a world average value of about 5.6°C per 1000m.

The seasonal variation of temperature lapse rates is illustrated here by data for different seasons and different elevations on Leigong-Shan, a mountain in Guizhou, in the humid-subtropical climate of southern China (Leigongshan Committee 1989). Note that the vertical differences (lapse rates) are greatest in summer (July), close to the world TLR average, suggesting greater heating in the lowland than at higher elevation, where atmospheric mixing is faster and more complete. The vertical difference is generally lower in winter (January), which may result from cold-air drainage (and an inversion layer below the lowest level shown here, 640m). Note also that the lapse rates here are consistently and significantly higher at mid-elevation than above and below (close to the world-average TLR in all seasons).

Table 10: Monthly Temperature Lapse Rates for Different Sides of Mt. Leigong-Shan (Guizhou)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
West side	.42	.47	.44	.43	.46	.50	.52	.50	.50	.42	.41	.40	0.46
East side	.40	.44	.40	.41	.44	.47	.52	.52	.45	.41	.40	.40	0.44
North side	.38	.42	.41	.43	.44	.50	.56	.55	.52	.41	.39	.38	0.45
South side	.44	.51	.45	.45	.47	.50	.50	.48	.51	.44	.44	.44	0.45

Variation in temperature lapse rates (difference per vertical distance upward) is shown here for different sides (and by month) on Leigong-Shan, a mountain in Guizhou, in humid-subtropical southern China (see also Table 9; data from Leigongshan Committee 1989). The vertical temperature differences (lapse rates) are again consistently greater in summer than in winter, as in Table 9 for different elevations. Note here that both lapse-rate extremes (greatest and smallest vertical temperature differences, shown in boldface) occur on the north side of the mountain. The differences, though, are relatively small, especially the yearly average. Note also that all lapse rates are significantly less than the world-average value of 6.4° C per 1000m for the free atmosphere but are generally close to the world average Terrestrial Lapse Rate of 5.6° C per 1000m in summer.

Precipitation Patterns Upward

Orographic effects on precipitation are well known, if not easily or often quantified. Due to orographic lifting and adiabatic cooling, not only humidity but also cloudiness and precipitation amounts usually increase upward from a mountain base, both on the windward and leeward sides, up to a characteristic cloud level (if there is one). If the mountains are high enough, precipitation may decrease above that cloud level. This general vertical pattern for precipitation amounts, postulated as doubling for each 1000m upward while below cloud level and halving for each 1000m above, was incorporated into the model POLATE used (see above) for estimating the temperature and precipitation on Jianfeng (Hainan), based on triangulation from nearby sites with measured data (Box, unpublished).

Table 11: Temperature and Precipitation at Different Elevations around Lei-Shan (Guizhou)

Elevation	Ty	Tmmin	Tabmin	Py	Years
840m	15.4°	7.9°	-8.9°	1337mm	20
641m	15.7°	8.1°	-10.3°	1133mm	22
500m	16.7°	8.2°	-8.1°	1226mm	22
286m	18.1°	8.1°	-5.8°	1211mm	27

Temperatures and precipitation are shown for different elevations on Lei-Shan in southern China (Guizhou), as measured over 20-27 years (from Leigongshan Committee 1989). Note that mean annual temperature (Ty) decreases gradually upward, but mean low temperature in winter (Tmmin) decreases less, perhaps due to the effect of the large monsoonal pool of cold air over interior Asia in winter. Absolute minimum temperature (Tabmin) is lowest at mid-elevation (641m), suggesting the effect of an inversion layer, with less cold air above. (Note, though, that the absolute minimum has been tracked for less than 30 years, so the true absolute minimum that could occur is probably lower.)

Precipitation is affected more by summer conditions and does not seem to vary greatly over the < 600m shown here. There may be a general tendency for increase, but it starts above 500m.

Abbreviations: Ty = mean annual temperature; Py = average annual precipitation; Tmmin = mean minimum temperature in the coldest month; Tabmin = absolute minimum temperature (lowest ever measured – but here for < 30 years).

Some insight into vertical precipitation patterns may be given by the data in Table 11, from Lei-Shan mountain in Guizhou (albeit for only 20 years). Annual precipitation is shown as greatest at 840m and lowest at 641m, which does not immediately suggest any identifiable cause. There may be no permanent cloud layer, but 840m is approaching the level (above mountain base) where clouds may develop frequently. Precipitation may depend significantly on aspect, which was not given in the data. The effect of aspect and other geographical features on mountain precipitation was studied by Basist (1989), who tried to relate precipitation amounts, by multivariate statistical analysis, to such factors as mountain orientation relative to prevailing wind direction and to the fetch of the winds, i.e. the unobstructed distance from mountains to the source of wet air (truncated at 100km). The results were not conclusive, due partly to difficulty in measuring the factors accurately. This study does, however, suggest a perspective to be pursued.

Vertical precipitation patterns in various tropical mountain areas were summarized by Lauer [62; see also Fig. 31 in 65] and suggest four characteristics:

- an increase upward to some maximum level, usually 1000-2000m above sea level;
- increasing height of this maximum level in drier climates;
- greater precipitation increase at the maximum level in some wetter climates; and
- a secondary maximum in some mountains, well above the main cloud level.

Lauer also showed complete vertical precipitation profiles for two tropical mountain areas, both rising from relatively low elevation. One is on the eastern slopes of the Mexican highlands [61] and the other on the eastern slope of the eastern Andes cordillera in Ecuador [64]. The vertical precipitation patterns [summarized in 65] were juxtaposed with the altitudes of maximum cloud development, the fog belt, and the vertical vegetation belts. Precipitation is shown as greatest at about 1000-1500m, just below the level of maximum cloud development. Above the cloud level, precipitation decreases back to lowland levels (or less), but it then increases again to a modest

secondary maximum at about 3500-3800m (decreasing again above that). Since rainfall increases downward within clouds by collision-coalescence of raindrops, it is reasonable that the precipitation maximum would be just below cloud base. Precipitation should decrease below that maximum level due to re-evaporation of some rain. The secondary precipitation peak is explained as follows: after the rain has fallen in the area of the precipitation maximum, there is a significant saturation deficit, which is reduced again upward by convection from the cloud level, until a second condensation level is reached [65 - p. 107].

Perhaps even more importantly, there have been suggestions in recent decades of a general rise in the height of the cloud base and a consequent drying out of montane forests, especially in the humid tropics, with adverse consequences for mountain ecosystems and their biodiversity [e.g. 79, 58, 66]. This problem has been studied especially around the Monte Verde ecological station in Costa Rica, where a rise in cloud base and consequent lengthening of seasonal dryness were attributed first to rising sea-surface temperatures in the Caribbean [80, 91; cf 28, 11] and then to deforestation in the adjacent Caribbean lowlands [67, 75]. This rise in cloud base may, however, be a more general phenomenon, as suggested by 30 years of rising average cloud base (by more than 100m) in the northern Appalachian Mountains of eastern North America [83, 25]. Global warming will exacerbate this problem, since warmer air must rise further before condensation can occur.

And finally the Joker – Temperature Inversions

In some cases, temperatures may increase upward rather than decrease. Textbooks give at least five ways that such temperature inversions can arise: from radiative surface cooling at night, from cold-air drainage along slopes, by warmer air overriding colder air (warm front), by advection of warm air over a cold surface, and by adiabatic warming in a subsiding air mass [24; cf 1, 2]. In mountainous terrain, the most important of these mechanisms is cold-air drainage into valleys, including upland valleys in mountains and large river valleys between broad uplands, especially in polar, boreal and temperate continental areas with long cold winters (Siberia, China and Tibet, Middle Asia, Canada, Alaska, and even some temperate-zone mountains). In the western USA, the widespread occurrence of treeless valleys (as in Photo 5) has been attributed by some to cold-air drainage, which shortens the effective growing season in valleys [e.g. 12]. It is difficult to study such temperature inversions, though, because, as in most areas, the meteorological stations are not in the uplands but rather in the valley bottoms, where it is the coldest.

The occurrence of inversion layers is well understood, but their depth and the magnitude of their effects on vegetation are not always appreciated. One example may suffice, from mountains in northwestern China, where in 2006 we (Kazuo Fujiwara, local botanists, and I) were studying the forest vegetation at 1800-1900m in the Xiaolong-shan (34.37°N, 106.21°E), above Tianshui, which is situated at about 1175m in the Wei river valley in southeastern Gansu province. Mean monthly temperature and precipitation for Tianshui, the closest nearby meteorological station (34.58°N, 105.75°E), measured over the years 1935-1988, are shown in Table 12. Mean annual temperature was 10.9°C, and mean minimum temperature in January was given as -6.7°C; average annual precipitation was 533 mm. There are two months with mean temperature below 0°C; absolute minimum temperature was estimated by triangulation (program POLATE) as about -17.5°C and given in the Russian source as -17.4°C (years of measurement

not given). According to the normal rule-of-thumb [e.g. 114], this extreme temperature below about -15°C , even if for only short duration, would preclude evergreen broad-leaved woody plants, even when adequate water is available.



Photo 5: Treeless Valley with Conifers on Adjacent Slopes

Growing seasons may be shorter in valleys, where cold air remains longer in spring and appears earlier in autumn than on adjacent slopes. This photo shows a conifer forest with aspen pioneers (*Populus tremuloides*) near the North Rim of the Grand Canyon in northern Arizona. The forest boundary may be sharpened somewhat by fire or grazing, but the phenomenon of treeless valleys, with relatively flat bottoms, was attributed to cold-air drainage already by Billings (1954). Such treeless valleys are common throughout the mountainous western USA and Canada.

Temperatures at the study site (1870m) were extrapolated from the valley data, using a terrestrial lapse rate of $5.5^{\circ}\text{C}/1000\text{m}$, yielding a mean annual temperature of 7.1°C , mean minimum of about -10.8°C , and an absolute minimum of -21.5°C . Average annual precipitation was estimated (program POLATE) as 695 mm. The estimated mean monthly values are shown in Table 12. These projections suggest that, at the study site, at least 3 months should have mean temperature below 0°C – but the projections are based on data from the valley inversion layer. Even with allowances for this, though, would mean winter temperatures at 1800m not still be too cold for evergreen broad-leaved trees?

Since the water balance was much ameliorated on the upper slopes, there was forest at 1800m, dominated by deciduous oaks, mainly the warm-temperate deciduous species *Quercus liaotungensis* and *Qu. aliena* typical of northern China [35, 94, 117]. Growing among them, however, was also a moderately large oak tree with leaves that, in comparison with the other oaks around, were clearly:

- thicker, perhaps from the previous growing season;
- significantly darker; and
- apparently evergreen.

The occurrence of even one evergreen oak generally means that the rest of a [deciduous] forest is secondary. In fact the evergreen oak was identified as *Qu. spinosa*, which according to the “Flora of China” [115, vol. 4] ranges naturally from Jiangxi and Fujian to Yunnan, and occurs in Taiwan and Tibet.

Table 12: Base and Estimated Upland Climate on Xiaolong-Shan, near Tianshui (northwestern China)

Base conditions (1174 m): Mean monthly temperature and precipitation at Tianshui (from www.worldclimate.com):

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
T:	-2.3	0.7	6.4	12.5	17.0	20.7	22.8	21.7	16.5	11.2	4.7	-1.2	10.9° C
P:	4	6	16	40	58	74	92	93	91	42	14	3	533 mm

Mean minimum temperature in January: -6.7°C, absolute minimum: -17.4°C

(at 1143m; from the Russian archives at www.climatebase.ru, years not given; retrieved 2013)

Study site (1870 m): Projected mean monthly conditions, based on lowland conditions and a terrestrial lapse rate of about 5.5°C per 1000 m:

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
T:	-6.1	-3.1	2.6	8.7	13.2	16.9	19.0	17.9	12.7	7.4	0.9	-5.0	7.1° C
P:	5	8	21	52	76	96	120	121	119	55	18	4	695 mm

Estimated mean minimum T in January: -10.8°C, estimated absolute minimum: -21.5°C

Tianshui is located at 34.58°N, 105.75°E, in southeastern Gansu province of northwestern China

(elevation 1174m). A forest study site was located at 1870m in the Xiaolong-Shan mountains above Tianshui. The projection of climatic conditions upward to the study site was based on a Terrestrial Lapse Rate of about 5.5°C per 1000m for China and a doubling of precipitation (below the cloud belt) for every 2000m, as incorporated in program POLATE. An even smaller lapse rate could be used (cf Tables 10-12), as for other mountains in China, but would still suggest much lower temperatures above than in the valley below.

The occurrence, at Tianshui, of mean temperature below freezing for two winter months, plus the absolute minimum of -17.4°C, mean that evergreen broad-leaved trees would not be able to grow there, even if precipitation were sufficient (which it is not). Such trees would also be precluded in the presumably colder mountains above. Even so, at the study site almost 700m above, a large individual *Quercus* tree was found, which had thick, significantly darker leaves that appeared to be evergreen. It was identified as *Qu. spinosa*. Such an evergreen tree could only occur there if the temperature is not lower but rather at least 2-3°C higher than in the valley below. This suggests a significant, vertically thick inversion layer and surprisingly warmer conditions above.

We will never have the opportunity to return to that location in winter to measure the temperatures, even for short durations. So we encourage colleagues to make the relevant winter temperature measurements at higher elevations, in mountains everywhere, and to share them with all of us. This paradox has direct, significant implications for vegetation-climate studies in complex terrain and in high-latitude regions, as well as for broad-scale vegetation modeling and mapping in general. The depth of inversion layers in valleys can be large, reaching 1000m or more, as suggested by Lauer and Rafiqpoor [65, see Fig. 35]. If one could include an inversion layer of appropriate depth, temperatures could perhaps be estimated that would not contradict observed vegetation – but how many broad-scale models consider inversions?

Conclusion

What does the above mean for the delimitation of uplands? It does not answer many questions but does, hopefully, provide a more or less unified geographic framework for upland zonation. As a start, it appears we can say that uplands start:

- somewhere above the maximum seasonal height of an inversion layer, if it occurs (unless the upward warming crosses some enabling growth threshold);
- below cloud base; and
- where vegetation changes.

Also, uplands should start where the length of any dry season is decreased significantly. As shown above, air at 50% relative humidity would cool adiabatically to 100%, with cloud formation, after a rise of only about 1000m. So, as a crude, round-number, general rule-of-thumb, one can often expect cloud base to occur about 1000m above the base of the mountains. If the base is in the lowlands, near sea level, then uplands might “start” somewhere near but below 1000m. Uplands can start below cloud base because saturation is not required in order to reduce evaporative demands to the point of ameliorating daily or seasonal drought significantly, which may be reflected in vegetation type.

Inclusion of uplands and belts in classification systems will vary with the base climatic zone, since thresholds involve mainly temperature in wet climates and at high latitudes but usually moisture balance elsewhere. The nature of apparent main thresholds and some possible climatic measures are summarized in Table 13. In polar climates, uplands start where tundra (with vascular plants) disappears. In boreal climates, uplands (which are immediately alpine) start with the treeline krummholz, which may not be far above the lowland. Temperature correlates have been suggested for both of these. For climates with no dry season (and thus forest vegetation), upland delimitation generally involves one of two temperature-based limits:

- warm periods long enough for lowland trees to complete their life cycles, usually about four months above 10°C for temperate-zone trees; or
- temperatures low enough (tropics) to inhibit photosynthesis (especially difficult to assess).

For climates involving water-balance thresholds, delimiters based on the length of the dry and wet periods are probably easier to identify and to quantify than is water balance.

Table 13: Nature of Upland Thresholds and Possible Climatic Measures

		Nature of Upland Thresholds	Possible Climatic Measures
Wet climates	I	metabolic inhibition by low temperatures	[needs research]
	Ve, Vm	low temperature means or extremes	$T_{min} < 0^{\circ}\text{C}$, $Tabmin < -15^{\circ}\text{C}$
	VI	length of warm period	4 months $\geq 10^{\circ}\text{C}$
	VIII	length of warm period	1 month $\geq 10^{\circ}\text{C}$
Dry climates	III, VIIa	appearance (upward) of a sufficiently wet period	$L_{GS} \geq 1$ month
Wet-Dry climates	II, IV, VII	length of wet (warm) period	$L_{GS} > \text{lowland } L_{GS}$ by 1-2 months
Polar climate	IX	warmth threshold (significantly above 0°C)	$T_{max} \geq 3^{\circ}\text{C}$

Abbreviations: T_{max} , T_{min} = mean temperature of warmest/coolest month, respectively; $Tabmin$ = absolute minimum temperature (lowest extreme ever); L_{GS} = length of a growing season (months)

The full sequence of altitudinal belts in mountains is generally fairly well identified and understood conceptually, in all the main climatic zones, but the belts are still recognized mainly by their vegetation rather than by climatic factors. For these upland belts, in summary, it appears that:

- climatic delimitation of the alpine may be possible, but
- other belts are more problematic, due to apparent water-balance limitation.

Some climatic data are available, but not nearly enough. One important need is for more climatic data (at least monthly temperature and precipitation) at different elevations and on different sides of mountains, in all climatic zones, in order to estimate lapse-rate variations more accurately. Detailed water budgets would also be useful – but with variables that can be calculated with available data, for modeling purposes. For the vegetation, one need is for better understanding of limiting factors in mountains, especially above tropical lowland rainforests. Much insight into vertical precipitation patterns and controls in tropical mountains was gained from the work of Lauer, but this work seems not to be well known outside German-speaking Europe.

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REGIUNILE ÎNALTE ȘI ZONAREA GLOBALĂ

(Rezumat)

Primele încercări de a regionaliza clima pe glob au început în anii 1800, bazate mai ales pe relația empirică cu peisajul natural. Mai apoi, conștientizarea circulației atmosferice globale a dus la apariția clasificărilor „genetice” și a conceptelor de zonare, bazate pe mecanismele atmosferice. A fost universal recunoscut faptul că zonarea urmărește modelul temperaturii, cu sezoane umede și uscate ca factori suplimentari. Toate aceste concepte au fost bazate pe modele din regiunile joase – dar ce sunt regiunile joase? Și cum ar putea fi extinsă sistematic zonarea bioclimatică până în regiunile înalte?

Regiunile înalte sunt zone unde peisajele naturale sunt semnificativ diferite de cele din regiunile joase. Însă, regiunile înalte pot fi delimitate și prin criterii măsurabile, cum sunt punctele topografice, limita de unde temperaturile încep să scadă, gradientii bruști de temperatură sau umiditate, care pot să includă atât umiditatea aerului cât și precipitațiile. În insule și de-a lungul coastelor, straturile atmosferice de inversie și norii persistenți, pot de asemenea să delimiteze regiuni înalte.

Accentuarea sezoanelor secetoase, în special, este proiectată aici prin schimbări pe 1 km și 2 km în acele tipuri de climă care au perioade secetoase. Se fac sugestii și de criterii în ceea ce privește temperatura și balanța hidrică pentru a identifica începutul zonei înalte și etajele altitudinale înalte, care sunt afectate de scăderi ale temperaturii și modele verticale de precipitații care nu sunt consecvente din punct de vedere geografic, sezonier sau al pantei. De asemenea, este demonstrată importanța inversiilor de temperatură, însă felul în care acestea se estimează și se includ în modele geografice la scară mică rămâne iluzoriu.

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