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MET O 3 TECHNICAL NOTE NO. 21

THE VARIATION OF SURFACE TEMPERATURE WITH ALTITUDE

by

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Contents

Summary

Altitudinal gradients of temperature along the ground are determined by the lapse of temperature in the free atmosphere as modified by the surface. The effects of the surface can be considered in 3 stages:-

i. The diurnal and seasonal modifications to the lapse rate in the free atmosphere; these depend upon latitude, climate, and the state of surface.

ii. The influence of topography, where it is important to distinguish between local and large-scale effects.

iii. Changes in climate and state of surface following the ground.

Although explicit consideration of these three factors does not necessarily provide a practical means of estimating surface lapse rates, it does enable the wide variety of observed values to be understood and placed in perspective.

1. Introduction

Altitude is one of the main causes of temperature variations on the surface of the earth. The change in temperature associated with a movement of several hundred kilometres in the horizontal may be accomplished by a change of only a few hundred metres in the vertical. Relatively few meteorological observations are made in mountainous districts and it is often necessary to estimate temperatures at a given location from observations made at a different altitude. As a result, it is important that differences of temperature introduced by altitude be well understood.

The variation of temperature with height in the troposphere is relatively simple. An upper limit to a decrease with altitude of 9.8°C per km is provided by the dry adiabatic lapse rate resulting from the mixing of air unsaturated with water vapour. The saturated adiabatic lapse rate represents the effect of mixing of cloudy air and varies with temperature according to the amount of moisture available for latent heat release. Values range from about half the dry adiabatic rate at low levels in the tropics to very close to the dry adiabatic rate near the tropopause, but 6°C per km is a typical value. Whenever subsidence occurs in the atmosphere, more stable conditions prevail with isothermal and inversion layers being produced.

The final result of these factors is that mean temperatures in the troposphere decrease with altitude in a fashion similar to the saturated adiabtic lapse rate. Exceptions occur near the presence of semi-permanent subsidence inversions, such as occur over the sub-tropical oceans. The influence of the surface of the earth also causes modifications to lapse rates in the free atmosphere to occur on diurnal and seasonal time scales. Nevertheless, a lapse rate of 6°C per km has become accepted as standard, and is commonly used for reducing temperatures to sea level.

The lapse rate of temperature over the ground, that is the variation with altitude of temperature recorded in a screen 1 to 2 metres above the ground, is not the same as that in the free atmosphere. It also depends on the effects of the surface, and these vary with latitude, climate, state of surface, local and large-scale topography. This large number of variables, and the interactions between them, produces a very wide range of lapse rates over the surface. The variations within the UK are illustrated by Smith (1975), who finds that lapse rates reach a seasonal maximum in spring on the northern Pennines in summer on the eastern slopes of the Welsh hills, and in autumn on their western slopes.

The variations of surface temperature in mountainous districts have been studied for over a century. Most of the investigations have been local and have taken place in the Alps, where the number of observations available is greater than elsewhere. The reports of these studies are mainly contained in the German literature, but the findings of many of them are reported by Geiger (1965). A very extensive review of the literature is provided by Barry (1981).

Much of the work on the variation of temperature with altitude has been empirical, and the identification of the physical factors responsible for the observed variations in lapse rates has been of secondary importance - in many papers the contributions of the relevant variables have not been recognised explicitly. To the casual reader, therefore, the wide range of lapse rates reported, and the often inadequate physical explanations offered, present a very confusing picture. The main aim of this paper is therefore to present a simple and convincing account of the main factors involved in the determination of surface lapse rates, with special emphasis on the effects of large scale topography. A general procedure for estimating surface lapse rates is suggested, but this is unlikely to be of direct practical value; it is offered mainly to act as a framework for a general understanding of the subject.

2. Geographical and seasonal variations of lapse rate in the free atmosphere

The continents respond to changes in solar radiation much more readily than the oceans so that winter lapse rates over the land are less than over the sea and decrease with latitude in response to the decreasing elevation of the sun. Over the oceans, however, lapse rates increase with latitude due to the advection of progressivel colder air from the land. In summer, lapse rates over the land are greater than over the sea and again decrease with latitude, but over the ocean latitudinal variations are slight. These changes are illustrated in Fig 1, which displays profiles of mean temperature determined from radio-sonde ascents for 3 land and 3 ocean stations. Variations over the sea are small compared to those over the land, where it can be seen that diurnal modifications to the atmosphere commonly extend up to about 1km, and seasonal effects to 2km. At OWS'N', at 31^{o}N, some evidence of the sub-tropical subsidence inversion is present between 850mb and 700mb. Geographical and seasonal variations in lapse rate between 850mb-700mb and 700mb-500mb have been mapped globally by Maejima (1977).

Land temperatures respond fairly rapidly to seasonal variations in solar radiation, but time is required to transmit these changes to the atmosphere and ocean. Thus while the seasonal variations in temperature of the sea surface and atmosphere (as represented by 1000-500mb thicknesses) are approximately in phase, they are about a month behind those occuring over the land. This has the effect of making the atmosphere over the land more unstable in the spring compared with the autumn. The differences, however, are small compared with those between winter and summer.

3. Forced ascent

The lapse rate following the surface of the earth is not the same as that in the free atmosphere; it also depends on the effects of the surface, and the most direct of these is that of forced ascent. When air is forced either up (or down) by the surface, its temperature changes either at the dry or saturated adiabatic rate, irrespective of that in the free atmosphere. The effect can be very important on individual occasions - fohn events, for instance, and when the melting of falling snow has produced an isothermal layer down to the surface. On these occasions, amounts of snow lying on the lowest ground may be small, but forced ascent will produce sub-zero temperatures on relatively high ground, leading to the accumulation of substantial amounts of snow or glaze at modest altitudes (see, for example, Pedgley, 1969).

For mean temperatures, however, with which this paper is mainly concerned, forced ascent is not as important as appears at first sight. Surface adiabatic lapse rates caused by forced ascent will differ most from those in the free atmosphere when the latter is stable. Under these conditions, very little forced ascent is likely, either because winds are very light or airflow is occurring around, rather than over, the mountains. Substantial forced ascent is only likely when the lapse rates of temperature in the free atmosphere are close to adiabatic values, and hence when the modifying effects are small. Forced ascent and descent may, however, be responsible for modifying lapse rates by affecting the distribution of cloud and rain, but this is considered later under climatic effects.

4. Local and large scale topography

Differences between lapse rates following the surface and those in the free atmosphere are mainly due to the effect of radiation at the surface, and this varies on a diurnal and seasonal time scale. The extent to which surface lapse rates are affected depends on the topography, and in this respect it is important to distinguish between the effects of local and large-scale topography. The term 'local topography' is used to describe the differences between hills and valleys etc while the term 'large-scale topography' is used to refer to differences in altitude averaged over large areas, eg to distinguish between plateaux and plains.

Local topography is responsible for introducing differences in the lapse rate between maximum and minimum temperatures. A low level station is usually in a valley, while a high level station is commonly on a summit. The diurnal range in a valley will clearly be greater than that on a summit, and so the lapse of maximum temperatures must be greater than that of minima. Suppose, however, two sites of similar local topography are considered, one on a plain and the other on a plateau. The diurnal range of temperature at the two stations would be similar, and so the lapse rates of maximum and minimum temperatures would be the same.

The effects of local topography are not restricted to those affecting lapses of maximum and minimum temperature. The lapse of mean temperature in the diurnally modified atmosphere is often not the same as that in the seasonally modified atmosphere above it. In daytime in winter, the nocturnal inversion is not properly broken down and the diurnally modified atmosphere is more stable than the overlying layers. The extent of this modification will depend on the local topography and will be greater over a valley than over a plain. The phenomenon may be regarded as the development of a seasonal modification of the 'enclosed' atmospheres in valleys but, because it is a function of local topography, it is for convenience regarded as part of the 'diurnally' modified atmosphere. The seasonally modified atmosphere is intended to refer to the 'free' atmosphere above a level surface rather than the 'enclosed' atmosphere in a valley.

When considering altitudinal gradients of temperature measured along the ground, it is helpful to refer to 3 topographic models, and these are illustrated in fig. 2. The first relates to local, and the second and third to large-scale, topography. The differences between the models are illustrated using a schematic temperature profile appropriate to a clear winters night. Well above the surface, a lapse of 6°C per km is assumed; below this the seasonally and diurnally modified atmospheres are represented by isothermal and inversion layers respectively.

The first model is of a general lowland with isolated mountains projecting into the free atmosphere. The variations in temperature on the surface will then be similar to those observed in the diurnally and seasonally modified atmosphere. The second model is that of a smaller-scale plateau which projects into the seasonally modifed atmosphere without extending it vertically. Lapse rates across the surface will then be similar to those in the seasonally (but not diurnally) modified free atmosphere. The third model is that of a plateau so extensive that the seasonally modified atmosphere extends as far above the plateau as above the lowlands. Differences in temperature between the plateau and those that would have been observed at sea level are then determined by the lapse in the un-modified atmosphere, ie around 6°C per km. These last two models apply to gently sloping, as well as level, plateaux.

From a comparison of temperatures at sea level with those at an altitude A corresponding to the height of the plateau in models 2 and 3, fig. 2 shows that for a winters night,

i. on an isolated mountain, temperature increases with altitude (the diurnally and seasonally modified lapse rates are relevant),

ii. on a limited plateau, temperatures are independent of altitude (the seasonally modified lapse rate is relevant),

iii. on an extensive plateau, temperatures decrease with altitude as the plateau is raised or lowered (the standard lapse rate of 6°C per km is relevant).

A plateau so extensive as to build a seasonally modified atmosphere of the same depth as that which would occur over a comparable area of lowland (as in model 3) probably does not exist on earth. The seasonally modified atmosphere, however, will always extend to greater altitudes over high ground than over lowland, so most areas of high ground will have characteristics intermediate between models 2 and 3.

A generalised model of the effects of large scale and local topography is depicted in fig. 3. A large area of high ground will raise the height of the seasonally modified atmosphere, and to this extent may be regarded as belonging to the extensive plateau model of topography. Temperatures will decrease by around 6°C per km up to the height by which the seasonally modified atmosphere has been raised. The remainder of the high ground may be regarded as projecting into the seasonally modified atmosphere, and lapse rates will be determined by these seasonally modified lapse rates in the free atmosphere. By averaging height over a certain area, a smoothly varying surface can be obtained and used to define the local topography. This is needed to introduce different lapse rates for maximum and minimum temperatures, and also to identify valleys which may contain seasonally modified 'enclosed' atmospheres.

The diurnal and seasonal modifications to lapse rates in the atmosphere are represented in Fig. 2 as being linear. In reality, of course, they will be non-linear, especially in winter, when lapse rates in the free atmosphere will change gradually from negative (ie inversion) near the surface to close to the standard value of 6°C per km at higher levels. Thus on a gently sloping plateau conforming to model 2, for instance, increasing altitude would be associated with an increase in temperature at the lower levels which would be gradually retarded and eventually reversed at higher elevations.

5. Variation of radiation with altitude

The thinning of the atmosphere with increasing altitude has the effect of increasing both incoming and outgoing radiation. As a consequence, for topographically similar sites, diurnal range will increase slowly with altitude. As temperature decreases, however, outgoing radiation decreases according to Stefan's Law. This causes temperature to decrease slightly less rapidly with altitude over the land than in the atmosphere.

6. Coastal Effects

Large changes in the diurnal and seasonal modifications to lapse rates in the free atmosphere occur in the vicinity of coasts. As air passes from the sea to land or vice-versa, changes in the lapse rate near the surface occur very rapidly (within a few minutes). With the passage of time, these changes extend upward so that the diurnal modification is complete within a few hours. The seasonal modification is formed from the accumulated effect, over several days, of the diurnal changes.

Altitudinal gradients of temperature along the ground in coastal districts may be regarded as a combination of a horizontal coastal gradient combined with the vertical gradient that would have occured inland. The horizontal coastal gradient will enhance the altitudinal fall in winter, but oppose it in summer. As a result, altitudinal gradients on coastal slopes will be similar to those which occur in the free atmosphere over the sea.

7. Climatic effects

Surface modifications to lapse rates in the free atmosphere are strongly dependent on the radiation received by the surface and this is a function of both climate and latitude. Thus differences in lapse rates of maximum and minimum temperatures induced by local topography will be much greater where the climate is clear and calm than where it is cloudy and windy. The strength of the stable layer which forms over continents in winter will depend on latitude and the frequency of clear, calm conditions. The extent to which these modifications affect lapse rates over the ground depends, as has been shown, on the topography. Thus for the extensive plateau model, differences in temperature from those that would have been observed at sea level depend only on the lapse rate in the un-modified free atmosphere, and are independent of climate. Note, however, that such a model requires the build-up of a large depth of seasonally modified atmosphere above the plateau and this is much more likely to occur if winds are light than if they are strong. Thus the extent to which a given plateau approaches this model may vary seasonally in accordance with variations in the strength of wind.

Lapse rates following the ground will also depend on changes in climate, whether they be due to changes in the horizontal (eg distance from coast) or vertical (eg cloud cover). This will induce differences between lapse rates of maximum and minimum temperature, since the diurnal range in one type of climate will usually be different from that in another. Topography itself, of course, has a strong influence on climate. Thus lowlands on opposite sides of a mountain range often have differing degrees of continentality and exposure to given airmasses. Isolated mountains and relatively small areas of high ground often have a cloudier, wetter, and windier climate than the adjacent lowlands. Extensive plateaux,on the other hand, often experience a dry, radiation dominated climate. Mountain slopes on the windward edge of the plateaux often have a moist, cloudy climate due to the effects of forced ascent, but the air has lost most of its moisture by the time it arrives on the plateau.

Changes in the state of surface (vegetation and soil moisture) will also affect surface lapse rates through the proportion of radiation which is used to provide evaporation rather than surface heating, and there is an interaction with climate here. Thus, in summer, lapse rates in districts where cloud and rain increase with altitude will be greater than those where they decrease. The drier and sunnier location will not only receive more radiation, but the drier soil means that more of that energy will be available for surface heating.

8. Snow cover

The importance of the state of surface is also seen through the effects of snow cover which generally causes a reduction of temperature compared with that observed over exposed ground. Altitudinal gradients of surface temperature will be most affected when snow free terrain at low levels is replaced by a snow cover at higher altitudes. The effects are likely to be greatest during the period when the snow is melting; some guide to the start of this period is given by the date of maximum snow depth. Using data from Geiger (1965), table 1 shows that in the Austrian Alps, this ranges from January near sea level to June above 3000 metres.

The evolution of seasonal variations in the depth of snow cover are illustrated by considering two stations whose mean temperatures differ by, say, 4°C. During the winter, when temperatures at both stations remain below freezing, snow may be accumulating at the same rate at both stations. Suppose that the warmer station reaches its maximum snow depth in March. As spring advances, snow depth at this station decreases while that at the colder station still increases. By the time the snow depth at the colder station has reached its maximum, that at the warmer station may have completely disappeared.

Thus, during the early winter, altitudinal and latitudinal gradients of snow depth are relatively slight. As spring arrives, a steep gradient develops between those regions where the snow has disappeared and those where it has reached its maximum depth. As spring progresses, this boundary advances into regions of greater and greater snow depth, and the gradient of snow depth increases. By June, at high altitudes and high latitudes, snow depth will increase rapidly from zero to very high values. The snow in this transition zone will be melting rapidly and may be expected to exert a considerable influence on temperature. When the highest in a series of stations is located in this zone of melting snow, and a linear relation is fitted to temperature against height, a large value of the lapse rate is likely to result.

9. Case studies

9.1 The Alps

Some of the most extensive studies of mountain climate have taken place in the Alps, and some data given by Geiger (1965), and reproduced in Fig. 4, provide a good illustration of the effect of topography on the variation of temperature with altitude. Fig. 4 displays the variation of mean temperature in January and July for various altitudes in the Austrian Alps. In summer, there is a reasonably linear relation between temperature and altitude with mean lapse rates close to 6°C per km. In winter, however, an isothermal layer forms between 700 and 1400m. 700m represents the average height of main Alpine valleys while they are still enclosed by mountains. Above 1400m sites may be regarded as being on the open slopes of mountains projecting into the free atmosphere. Between 700m and 1400m sites may be regarded as being enclosed, with the atmosphere 'projecting' into the land. The observed variation of temperature with height is therefore very much a function of the average topographical exposure of a site at a particular altitude. The figures given for 1400m for example, would not apply to a valley at that level surrounded by high mountains.

Radio sonde ascents in the vicinity of the Alps show that, in July, temperatures in the free atmosphere are very similar to those displayed in Fig. 4. In January, however, sonde ascents indicate that at the surface, 850mb and 700mb temperatures are about 1°C higher than those presented in Fig. 4. The estimated profiles of mean January temperature in the atmosphere above a surface at 200m outside the Alps, and above a valley at 600m in the Alps, have been sketched in Fig. 4. It is evident that in the region of the Alps, located towards the western margin of a continent, very little seasonal modification has taken place in the atmosphere, except that the lapse above 1400m is rather less than 6°C per km. Substantial modifications to mean temperatures extend only to about 500m above the surface, and do not therefore reach beyond the range of diurnal modifications. The top of this layer, however, has been raised by about 500m by the Alps, and so for the lowest 500m, temperatures decrease at approximately 6°C per km in accordance with an extensive plateau model of topography. Thereafter the influence of local topography is dominant, and the effects of the 'duirnally' modified atmosphere, which includes a lowering of maximum as well as of minimum temperature in winter, represent the typical topographic exposure of a site at a given altitude.

It may be noted in passing that the depression of maximum temperature in Alpine valleys in winter is not solely a product of radiation; the melting of falling snow also makes a contribution. It is well known that when falling snow reaches the melting level, the energy extracted from the atmosphere by the melting can produce an isothermal layer close to 0 deg C, and the snow gradually extends to lower and lower altitudes; the process is described by Lumb (1983). What is less widely appreciated is that such a mechanism will operate far more efficiently in light than in strong winds. While the melting of snow alone would produce an isothermal layer, the turbulent mixing caused by strong winds has the effect of producing an adiabatic lapse rate. Thus, in strong winds, the energy extracted from the atmosphere will be the same as in light winds, but the loss of energy will be re-distributed vertically. With strong winds, there is also the possibility of warm advection at low levels, either from air which has been over the sea or drawn from outside the precipitation area. Low level winds in the region of the Alps are often light, and the enclosed nature of Alpine valleys hinders large-scale mixing and encourages the development of an isothermal layer in the valley whenever falling snow reaches the melting level.

9.2 Eastern slope of the Rockies

One of the best examples of a gently sloping plateau occurs to the east of the Rocky Mountains in North America. Table 2 presents mean temperatures in January and July at two sets of stations, the first, Calgary to Winnipeg, at around 50°N and the other, Denver to Topeka, at around 40°N. Although the precise figures for each station will be influenced by the local topography, the main factor is the large scale topography. In winter, temperatures increase with altitude at around 9°C per km at 50°N, while at 40°N they are approximately isothermal. Evidently, the second model of topography, in which the high ground projects into the seasonally modified atmosphere, is appropriate. In Fig. 1, the profile for The Pas, at 53°N, shows that temperatures increase by around 4°C in the lowest kilometre of the atmosphere. The seasonally modified atmosphere will be deeper at Winnipeg, near the continental interior, than at Calgary, and this accounts for the additional temperature difference between the two stations. At 40°N, the greater power of the sun prevents the formation of such large temperature inversions near the surface (see Oklahoma in fig. 1), and so isothermal conditions prevail. In summer, temperatures decrease with altitude at around 4°C per km at 50°N and 3°C per km at 40°N. Both these figures are less than the 'standard' rate of 6°C per km. At 40°N, these differences are caused by changes in climate, which becomes drier and sunnier as one moves West from Topeka towards Denver. The increased radiation and decreased evaporation at Denver permit higher surface temperatures than if the climate were the same as at Topeka, thereby decreasing the lapse rate.

9.3 The Himalayas and Tibet

The outstanding example of the effect of climate on lapse rates occurs in the region of the Himalayas. The Tibetan plateau is semi-arid all year, whereas the plains to the south pass dramatically from dry conditions in May to wet in July. Table 3 compares the lapses of maximum temperature between Leh and Lahore in the West and Lhasa and Patna in the East. The differences in temperature between the high and low level stations are almost halved between May and July as maximum temperatures on the lowland are depressed by cloud and rain.

Lapse rates of mean annual temperatures are considerably less than 6°C per km, mainly due to the greater radiation and smaller evaporation at high levels in summer. These effects are opposed by the higher latitudes of the elevated stations and, especially in winter at Leh, to greater exposure to cold air masses from the North. The effect of the Tibetan plateau as a high level heat source in summer is well known, and is discussed by Flohn (1953, 1968, 1974).

The variation of temperature following the ground in the Himalaya region behaves ~~varies~~ in a non-linear fashion. At low levels, maxima decrease rapidly as the plains are left for the hillsides, and cloud and rain increase. Then, as the Tibetan plateau is approached, temperatures rise as moisture and vegetation decrease. Maximum temperatures at Lhasa are actually higher than at Cherrapungi, 2,500m lower! These changes may be ascribed to horizontal gradients resulting from the change in climate, even though that change is largely caused by the topography in the first instance.

9.4 The British Isles

Good accounts of the variations of lapse rates in Britain are given by Manley (1970) and Harding (1978). Harding (1979) also gives a detailed discussion of the factors operating in the northern Pennines while Green and Harding (1980) analyse data in southern Norway to infer probable conditions in Scotland.

With the exception of the Abderdeenshire plateau, the relief of Britain approaches that of the isolated mountain model of topography, and this forms the basis of the following description. Over the south-eastern half of England, the variation of solar radiation causes pronounced seasonal and diurnal variations in the lapse rate. Increased cloud and wind cause these variations to diminish towards the North West. In winter, lapse rates over the seas to the West of Britain increase with latitude, and this, in combination with the increased frequency of continental airmasses towards the Southeast, causes lapse rates over Britain to decrease from north west to southeast. These differences are illustrated in Fig. 4 using data from Crawley and Stornoway, and OWS'I' and 'J'. At Crawley, values exceeding 10°C per km in summer are partly due to the lack of complete radiational screening of thermometers in a Stevenson screen.

According to Harding (1978) and Green and Harding (1980), the greatest lapse rates on Britains northern hills are observed in spring. This is partly due to the melting of snow on the summits of these hills. The seasonal variation of lapse rates over the sea is opposite to that over the land, and this causes lapse rates measured along the coastal slopes of hills to reach a minimum in June and a maximum in December. The lapse of minimum temperature is non-linear, and increases gradually from near isothermal conditions in the lowest layers over the south eastern half of England to close to 6°C per km above about 1km.

10. Conclusions

Lapse rates of mean temperature in the free atmosphere away from the influence of the surface are generally close to 6°C per km, although exceptions occur near semi-permanent subsidence inversions such as occur in the zone of trade winds. Diurnal and seasonal modifications to this normal rate are caused by the effect of the surface, and the extent of these modifications depend on the latitude, climate, and state of surface.

Lapse rates of temperature over the ground can be estimated in 3 main stages:-

i. Take account of the location.- The latitude, climate, and state of surface~~, These~~ will determine the diurnally and seasonally modified lapse rates in the atmosphere.

ii. Introduce the topography-part raises the seasonally modified atmosphere, part projects into the seasonally modified atmosphere, and part produces local effects.

iii. Allow for changes in climate and state of surface following the ground. These are best regarded as being due to changes in the horizontal, even if they were largely caused by orography in the first place.

In more detail, the recommended procedure for estimating altitudinal gradients along the ground is as follows:-

a. Obtain the seasonally modified lapse rates at various heights above the surface in the free atmosphere. These can be obtained from radio-sonde ascents and, since they vary only on a large scale, can be mapped. For summer, and for the 850-700mb and 700-500mb layers, this has already been done by Maejima (1977). Especially in winter, these lapse rates will be non-linear and will change as the surface is approached. In the zone of diurnal modification, however, lapse rates in the 'seasonally only' modified atmosphere should be regarded as a linear extrapolation of those immediately above the zone.

b. Estimate the height to which the seasonally modified atmosphere extends, and the amount by which it has been raised over areas of high ground. This may vary with season.

c. Assume a lapse rate of 6°C per km up to an elevation which equals the amount by which the seasonally modified atmosphere has been raised. Thereafter the lapse of temperature in the seasonally modified atmosphere should be used.

d. Define a surface (eg by averaging height over a large area) which enables variations in height to be shared between a 'sloping plateau' and local topography. This will enable diurnal modifications to the free atmosphere, and seasonal modifications to 'enclosed' atmospheres to be taken into account.

e. Allow for changes in climate and the state of surface. By reducing temperatures to sea level using steps (a) to (d), these changes should be revealed as horizontal gradients, even though they were mainly caused by the orography.

The steps outlined above represent a generalised procedure which can only be implemented after a good deal of local study and empiricism, and ~~will~~ are probably impossible to apply in any strictly objective and quantitative fashion ~~prove impractical~~. They are offered mainly to act as a framework for a general understanding of the subject~~, which should~~ and to enable diverse results from different parts of the world to be placed in perspective.

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FIG 1 - PROFILES OF MEAN ATMOSPHERIC TEMPERATURE

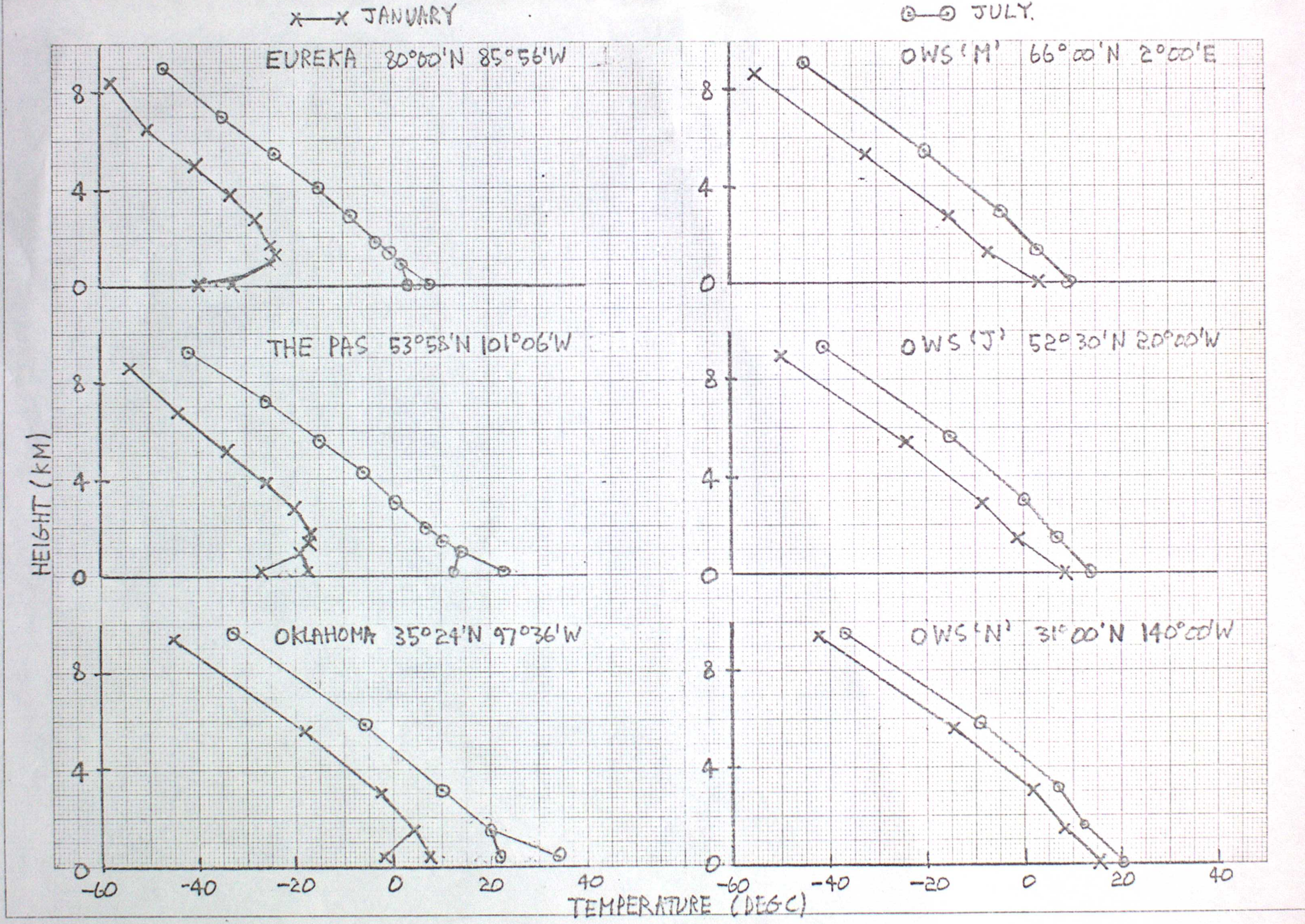

FIG 2 - SCHEMATIC REPRESENTATION OF TEMPERATURE PROFILES ASSOCIATED WITH 3 MODELS OF TOPOGRAPHY.

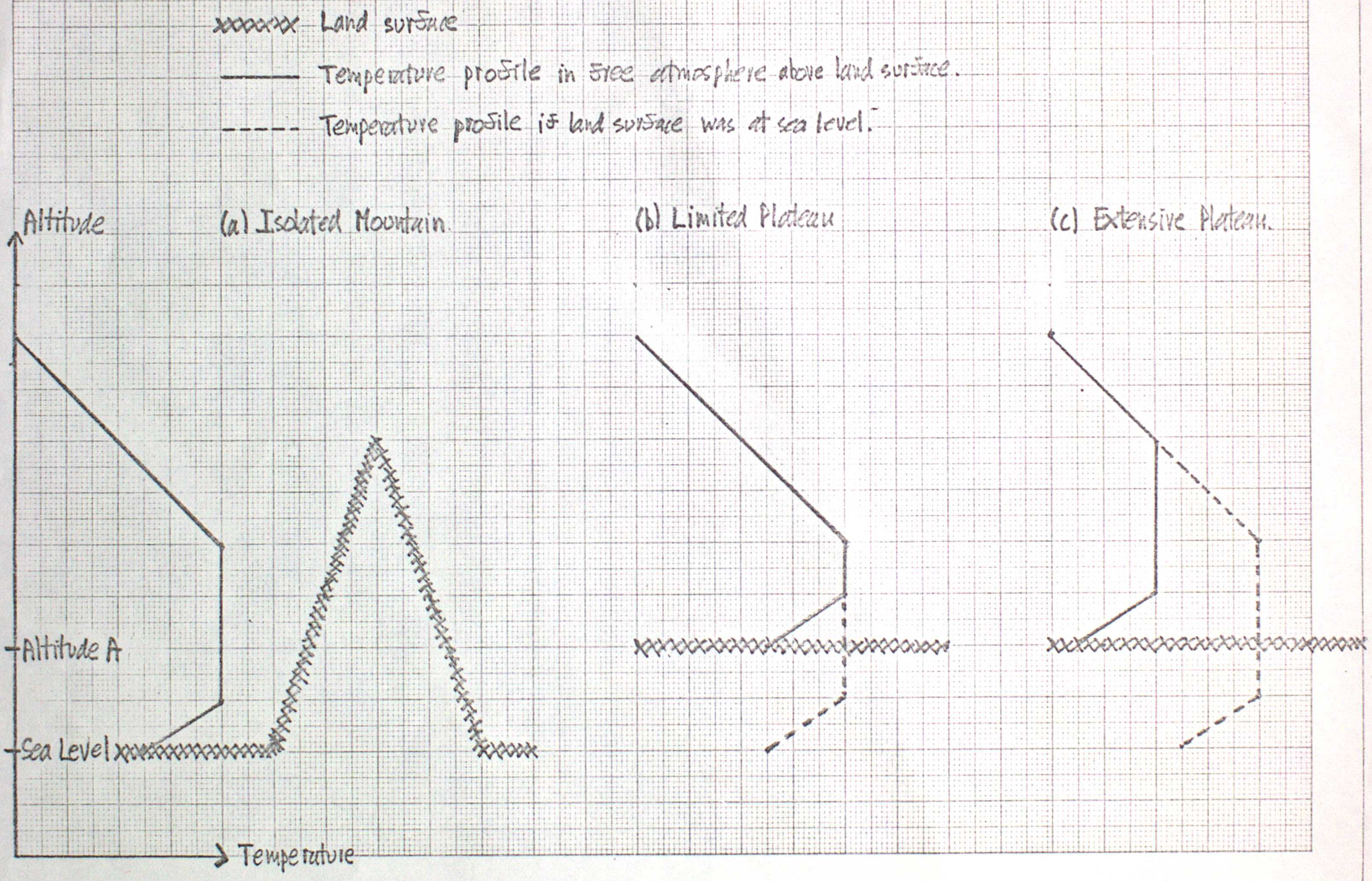

FIG 3 - COMPOSITE MODEL OF TOPOGRAPHY FOR ESTIMATING SURFACE TEMPERATURES.

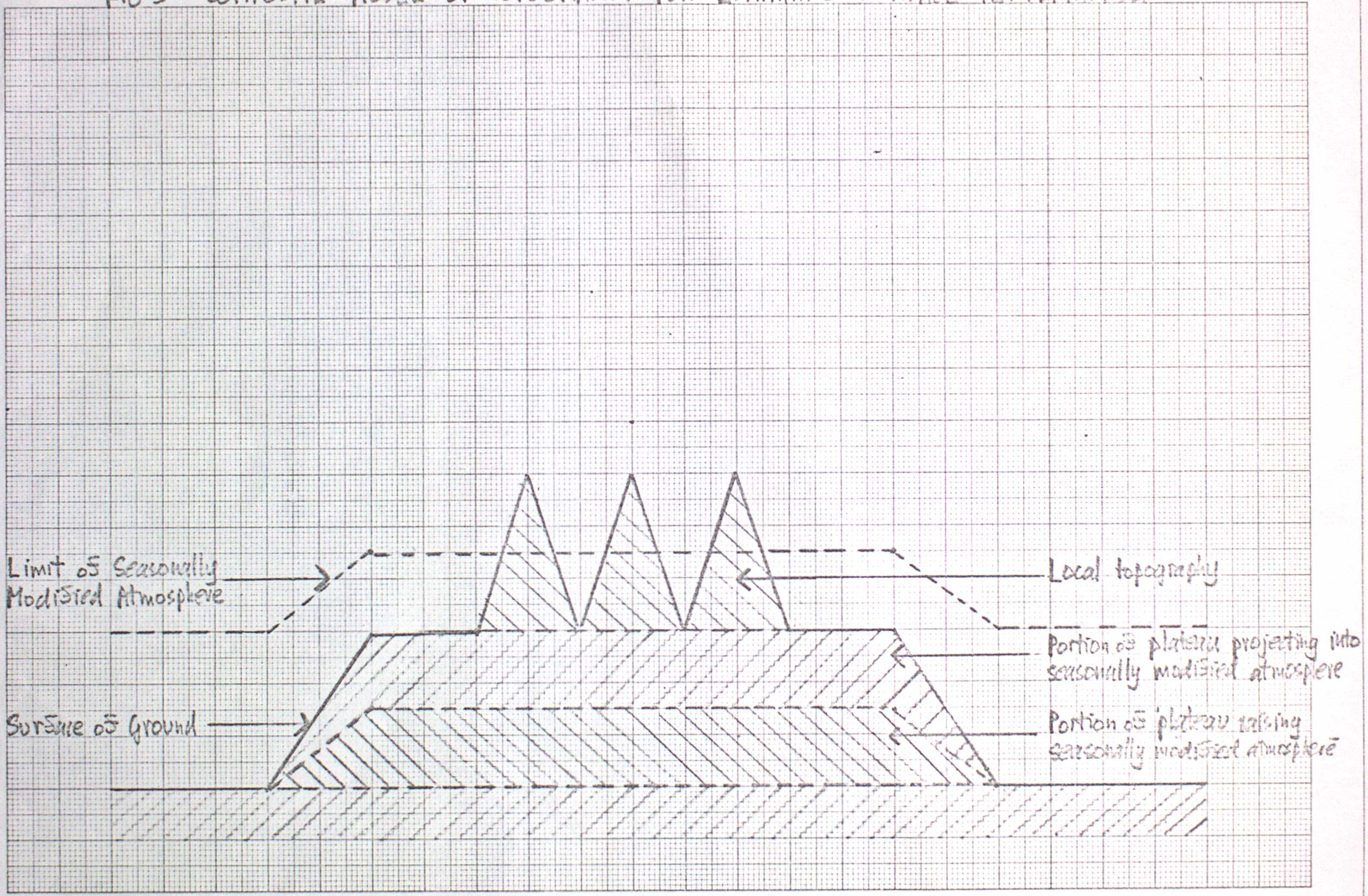

FIG 4 – VARIATION OF MEAN TEMPERATURE WITH ALTITUDE IN AUSTRIAN ALPS (AFTER GEIGER)

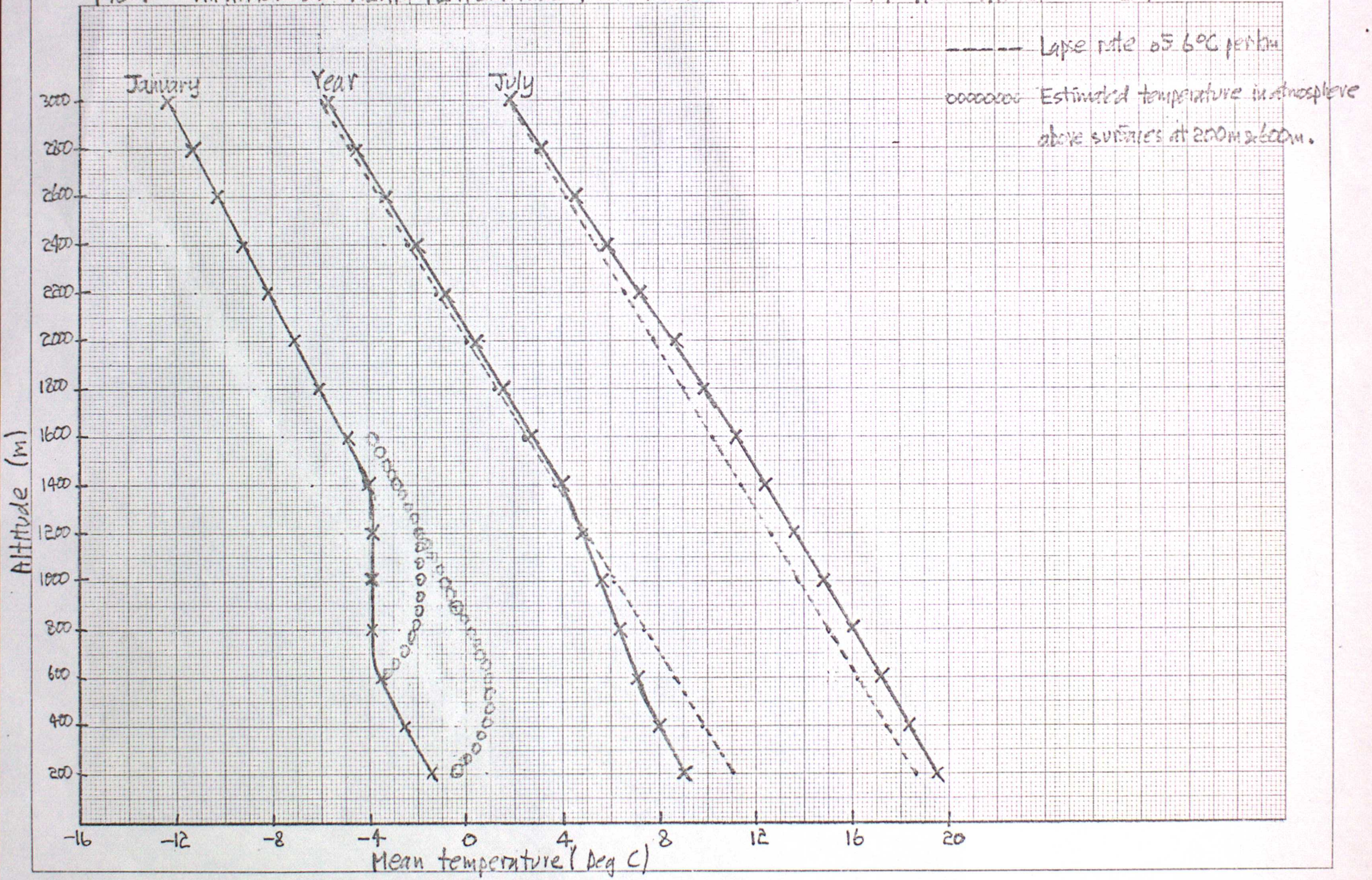

FIG 5 — MEAN LAPSE RATES IN THE FREE ATMOSPHERE IN THE VICINITY OF THE BRITISH ISLES FROM 1961-70.

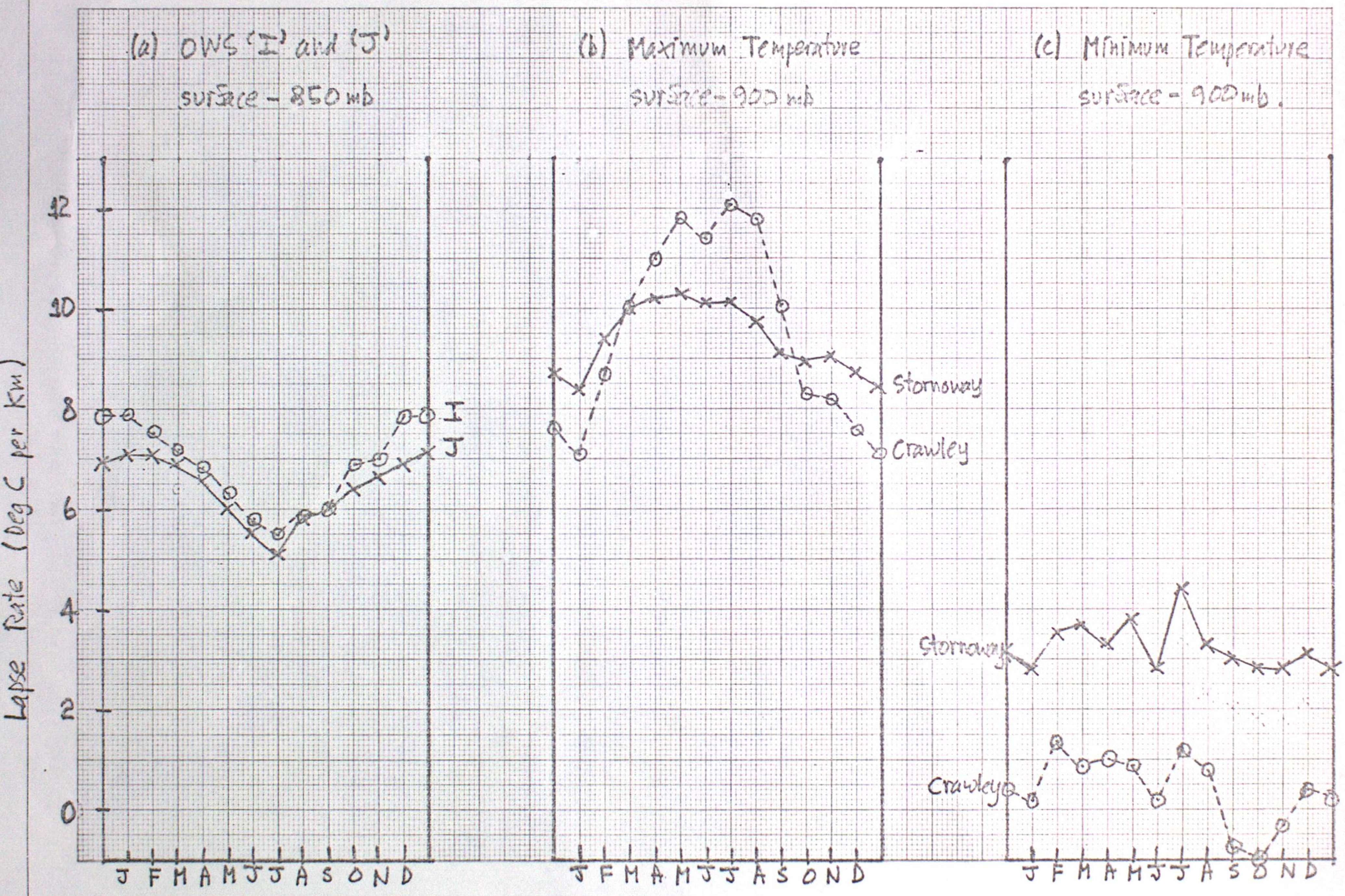

Table 1 - Date of maximum snow depth in the Austrian Alps (after Geiger)

Altitude (m)	200	600	1000	1400	1800	2200	2600	3000
Date	18 Jan	28 Jan	11 Feb	21 Feb	14 Mar	8 Apr	3 May	29 May

Table 2 - Mean temperatures on sloping plateau to East of Rocky Mountains.

Station	Altitude (m)	Jan	July
50°N			
Calgary	1079	−10·0	16·7
Swift Current	816	−13·1	18·9
Regina	574	−16·5	19·0
Winnipeg	240	−17·6	20·0
40°N			
Denver	1613	−2·0	22·7
Sterling	1201	−3·9	23·2
McCook	782	−3·0	25·2
Concordia	448	−2·6	26·8
Topeka	267	−1·8	26·6

Table 3 - Lapse Rate of Maximum Temperature in the Himalaya Region (deg C per km)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Leh-Lahore	6·6	6·6	6·4	6·6	7·3	6·4	3·9	3·7	4·6	6·1	6·1	6·2
Lhasa-Patna	4·4	4·6	5·7	6·0	5·2	3·4	2·6	2·6	3·1	4·1	4·3	4·1